Mass Movements in Great Britain

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Chapter 6

Mass-movement sites in Jurassic strata

INTRODUCTION

R.G. Cooper

The Jurassic strata in Britain (Figure 6.1) are susceptible to large-scale mass movements. Of the ten sites chosen for evidence of deep-seated slipping in Lower Jurassic strata, four have a substantial over-burden of non-Jurassic rocks that are also involved in the slides: in the Axmouth to Lyme Regis Undercliffs Nature Reserve (**Axmouth–Lyme Regis** GCR site) these are Cretaceous sediments, whereas at mass-movement sites on the **Trotternish Escarpment** on the Isle of Skye Paleocene lavas form a dramatic caprock. All of these sites are examples of the classic cases of failure where thick permeable strata overlay relatively impermeable argillaceous strata. The exception is at **Hallaig** on the Isle of Raasay, where the landslip is entirely in Jurassic strata, but at this site other factors may have had a greater influence than stratigraphy or lithology.

Figure 6.1 Areas of Jurassic strata (shaded) and the locations of the GCR sites described in the present chapter. The Storr and Quiraing lie within the larger **Trotternish Escarpment** GCR site.

The Lower Jurassic scarps through central England have been subject to cambering, valley-bulging and an attendant suite of features. These were first recognized in the Jurassic strata of the Northamptonshire Ironstone Field (Lapworth, 1911; Hollingworth *et al.*, 1944), but no Northamptonshire sites now provide good exposures of these features. Dip-and-fault structures can be seen in Jurassic strata at the **Entrance Cutting at Bath University** GCR site, and ridge-and-trough features can be seen in Jurassic strata at **Postlip Warren**, near Cheltenham. The site chosen to illustrate valley-bulge structures (**Rowlee Bridge**) has been described in Chapter 5, as it is in Carboniferous strata.

Illustrating the effects of the Lower Jurassic consisting largely of highly mobile clays, **Black Ven** is a landslide that is in the process of being actively degraded by mudslides, but episodically rejuvenated by other processes. It is formed in the Lower Jurassic argillaceous beds overlain by the Cretaceous Upper Greensand.

Large-scale superficial structures are also found in Upper Jurassic strata. Three sites, **Buckland's Windypit** and **Peak Scar** in North Yorkshire and **Blacknor Cliffs** on the Isle of Portland, give a broader impression of the types of mass movement in Jurassic strata. At Blacknor Cliffs clay extrusion and deep-seated settlement followed by rotation is typical. At Peak Scar the main mechanism is toppling.

The survey of landsliding (Jones and Lee, 1994) records 2236 examples in Jurassic strata, of which 55% are of unspecified type. The most common types are successive rotational slips (21% of those where the type is specified) and cambered/foundered strata (20%). The portmanteau classification type 'complex' accounts for 17% and single rotational slips for 13%. At the opposite extreme, 0.5% are topples, and no records of sags were obtained from Jurassic strata. The latter point is surprising: the Jurassic rocks in and around Dundry Hill, south of Bristol, are recorded by the British Geological Survey as 'foundered strata' because of the complexity of their superficial structures; this term suggests that the superficial structures are likely to include sags.

Thirty percent of the slides in Jurassic strata are recorded as having taken place in the Upper Lias, with the Inferior Oolite at 18%, the Middle Lias at 15% and the Lower Lias at 14%. Of the Upper Lias slides, 48% are of unspecified type, but of those where the type is specified, two types predominate: successive rotational slips (28%) and cambered slopes (24%) (Jones and Lee, 1994).

For convenience, the GCR sites selected are discussed here in two sections, covering the Lower Jurassic strata and the Upper Jurassic strata respectively.

MASS-MOVEMENT SITES IN LOWER JURASSIC STRATA

POSTLIP WARREN, GLOUCESTERSHIRE (SO 997 265)

R.G. Cooper

Introduction

Postlip Warren is an area at the top of the Cotswold cuesta, on Cleeve Hill, north-east of Cheltenham. Cleeve Hill is the highest part (300 m above OD) of the dissected Jurassic limestone escarpment, in a region of intra-Jurassic subsidence remarkable for its considerable thickness of Inferior Oolite (*c.* 107 m). The Inferior Oolite is composed mainly of limestones, with occasional sandy beds, for example the Harford Sands. The hill is deeply dissected by dendritic dry-valley systems that feed the dip-slope River Coln and the scarp-face river, the Isbourne. There are also two groups of anomalous troughs, one group occurring along the main scarp face above Prestbury and Southam, and another group truncating spurs near Postlip Warren (Figure 6.2).

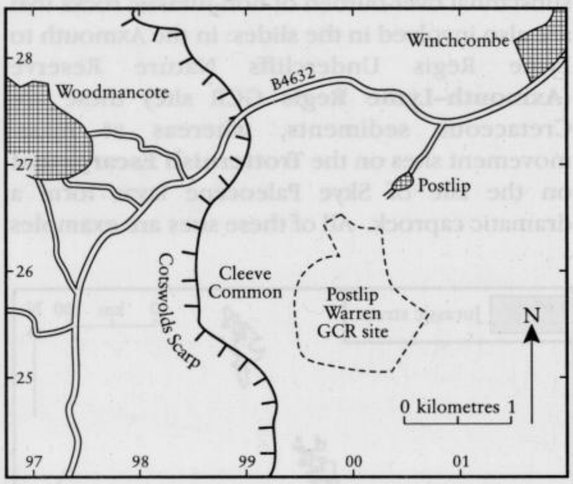


Figure 6.2 Location of the Postlip Warren GCR site.

Description

The troughs are dry and grass-covered, occurring at or near the crest of the escarpment, at heights exceeding 240 m above OD (Figure 6.3). The Postlip Warren group consists of three features truncating the spur between the Postlip

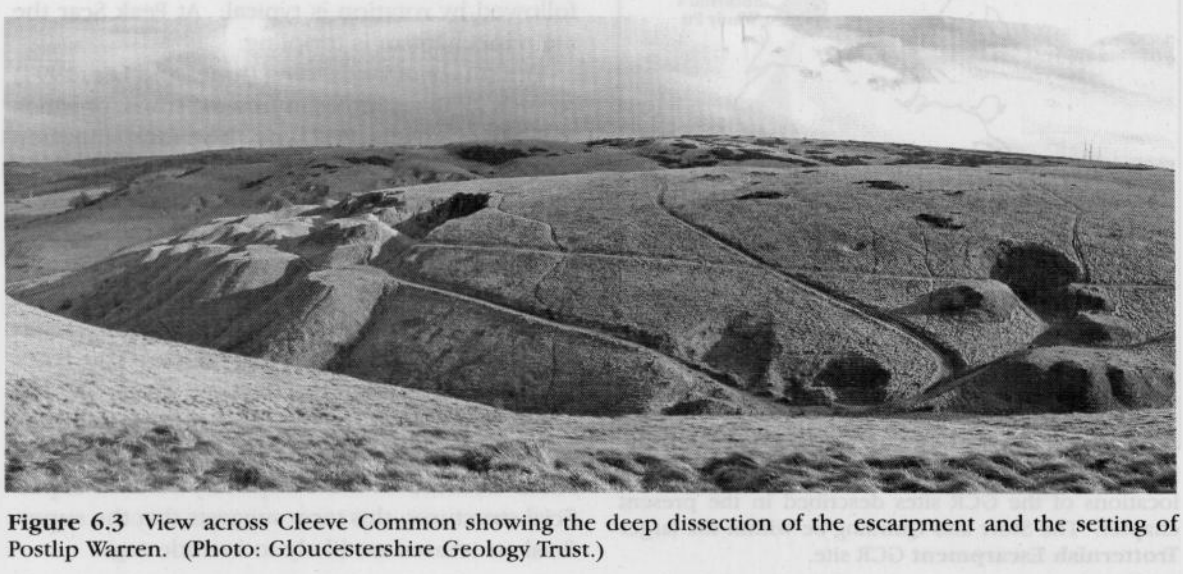


Figure 6.3 View across Cleeve Common showing the deep dissection of the escarpment and the setting of Postlip Warren. (Photo: Gloucestershire Geology Trust.)

and Corndean valleys. They are approximately parallel, and 12–15 m deep. They are aligned perpendicular to the major valleys. They have convex longitudinal profiles, with irregularities, rather than the normal concave longitudinal profile of most stream valleys. In places the maximum angle of the longitudinal profile approaches 10°. The bottoms of the troughs are broad and flat (Figures 6.4 and 6.5), and typically they have a width of 50–70 m from one break of slope to the other. From crest to crest, the widths are of the order of 150 m. They are characterized by asymmetry, with a tendency for the slopes on the plateau side to have a maximum steepness of 18°–21°, and those on the embayment side to stand at 10°–14.5° (Goudie and Hart, 1976). A further group of small depressions runs parallel to the Postlip Warren troughs. Most of them have a dominantly linear form, but they are in essence closed depressions. They contain a relatively deep fill of dark-brown clayey material, attain depths of 3–5 m and tend to follow the contours.

One of the main troughs ('Trough 3' of Goudie and Hart, 1976) contains more than 4.9 m of dark-brown clayey fill with oolitic fragments. The content of coarse oolitic material increases with depth.

Interpretation

The troughs are a distinctive type of landform because:

- they possess a constant asymmetry;
- they possess irregular or closed longitudinal profiles;
- they contain, in at least some cases, a deep, non-alluvial fill;
- they run parallel to the main relief trends;
- they truncate major drainage lines; and
- in some cases they rise where there is little or no catchment area.

Goudie and Hart (1976) argue that they are neither solutional nor glacial features. They point out that the deep fill and the closed nature of some of the features is consistent with a solutional origin, but that a solutional origin does not fully explain either the asymmetry of the cross-profiles of the troughs or the way in which the troughs run parallel to the main relief trends.

Another possible hypothesis for the origin of the troughs is that they are some type of glacial form. The up-and-down longitudinal profiles could have been formed by sub-glacial streams

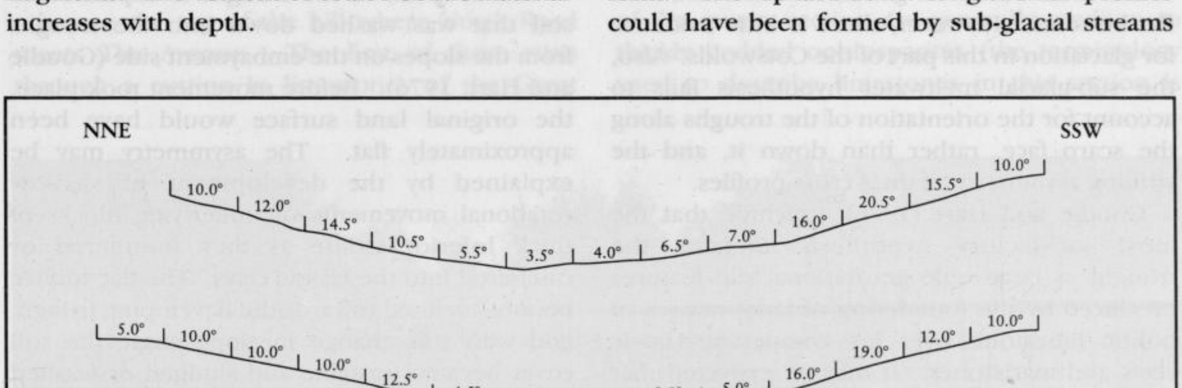


Figure 6.4 Representative slope profiles at Postlip Warren to show distribution of maximum angles. After Goudie and Hart (1976).

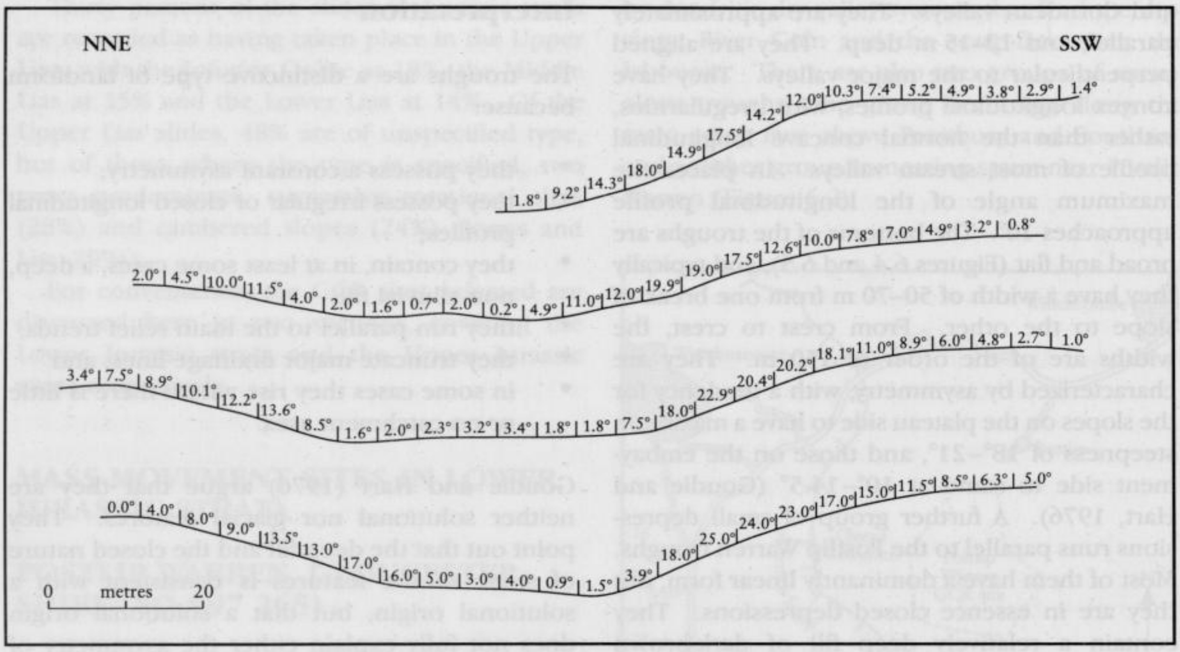


Figure 6.5 Representative profile at Postlip Warren to show broad and flat valley floors.

under hydrostatic pressure. The absence of a normal catchment for the troughs lends support to this idea. Many sub-glacial channels also have flat bottoms. However, there is little evidence for glaciation in this part of the Cotswolds. Also, the sub-glacial meltwater hypothesis fails to account for the orientation of the troughs along the scarp face, rather than down it, and the striking asymmetry of their cross-profiles.

Goudie and Hart (1976) conclude that the most satisfactory hypothesis explains the troughs as large-scale gravitational slip features produced by the foundering of large masses of oolitic limestone over less-competent Liassic clays and marlstones. It may be expected that the face along which slip took place would be relatively steep and that the opposite slope would be relatively gentle. Likewise it may be expected that the troughs would develop parallel to the edges of either the escarpment or embayments within it. Kellaway (1972) has suggested that over-steepening of the escarpment by ice coming down the Severn Vale is a possible contributing factor, though accelerated spring sapping or periglacial cambering could have similar effects. Small-scale superficial cambering is evident in many quarry sections on Cleeve Hill (notably at SO 987 272). Once the depressions were formed, water may have flowed along them under nival conditions (Beckinsale, 1970) and solutional activity may have accentuated initial irregularities.

The deep fill of the troughs is explained as soil that was washed down into the troughs from the slopes on the embayment side (Goudie and Hart, 1976). Before movement took place, the original land surface would have been approximately flat. The asymmetry may be explained by the development of shallow rotational movements of underlying blocks of thick Inferior Oolite as they foundered or cambered into the Liassic clays. The flat surface became inclined towards the developing trough, and with this change in slope angle the soil cover became unstable and sludged or washed down. On this hypothesis there need be no catchment area and the depressions might be expected to truncate the main dendritic dry-valley systems of the area.

The troughs bear comparison with the 'ridge-and-trough' features at Lower Slaughter in the north Cotswolds and are similar to features described generally in the literature as gulls, vents, grabens, dip-and-fault structures, rock labyrinths or camber crests (see Brunsden, 1996b).

Conclusions

Postlip Warren provides very clear physiographical evidence of large-scale gravitational slip processes. It exhibits the longest, deepest, and most-pronounced gravitational troughs in Great Britain, clearly displaying asymmetry and deep fill.

Although very little is known about this site, the dates or the mechanism involved – there is no sub-surface evidence – there is the opportunity to study the subject of material spreading at this site. There is both academic and economic importance in the subject because the forms imply that the whole hill may be in a 'residual strength' condition and might easily be re-activated by inappropriate civil engineering activity.

ENTRANCE CUTTING AT BATH UNIVERSITY, AVON (ST 767 645)

R.G. Cooper

Introduction

Bath University of Technology is sited on Bathampton Hill to the east of Bath at 204 m OD. It has two main roadway entrances, one 400 m down the hill on North Road, and the other on the top of the hill where North Road meets The Avenue. The first of these runs through a cutting in limestones of the Great Oolite Series (Lower Jurassic) (Figure 6.6). In the cutting, the rockface on the western and north-western side of the road shows very clear examples of dip-and-fault structure (Figures 6.7 and 6.8).

The City of Bath is situated in one of the most intense zones of landslipping in Great Britain. This is largely due to the presence of Lower Jurassic limestones overlying incompetent Lias Clays, in an area where the River Avon and its tributaries are deeply incised in steep-sided valleys (Kellaway and Taylor, 1968; Chandler *et al.*, 1976). The largest-scale movements took place during Pleistocene times. The Great Oolite is cambered on the western, northern and eastern slopes with downslope dips of up to 37° (Hawkins and Kellaway, 1971). Gulls that result from this cambering have been described by Hawkins (1977).

Description

Until 1967 the only outcrop at the site was an old quarry. The roadway to the university was then cut through the Great Oolite, revealing that the strata are in a disturbed state (Hawkins, 1977). The basal 2 m seen in the northern part of the quarry section are generally medium to thickly bedded oobiosparites (the terminology used to describe limestones in this section is

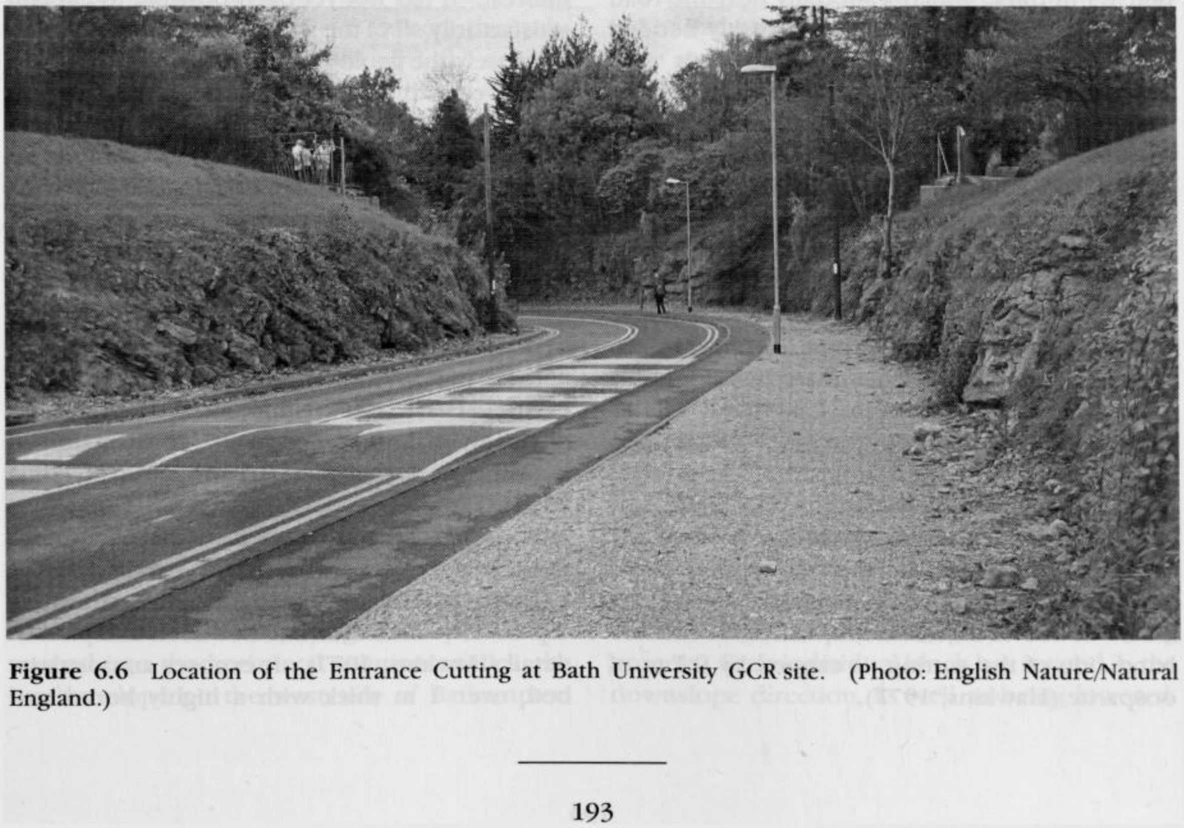


Figure 6.6 Location of the Entrance Cutting at Bath University GCR site. (Photo: English Nature/Natural England.)

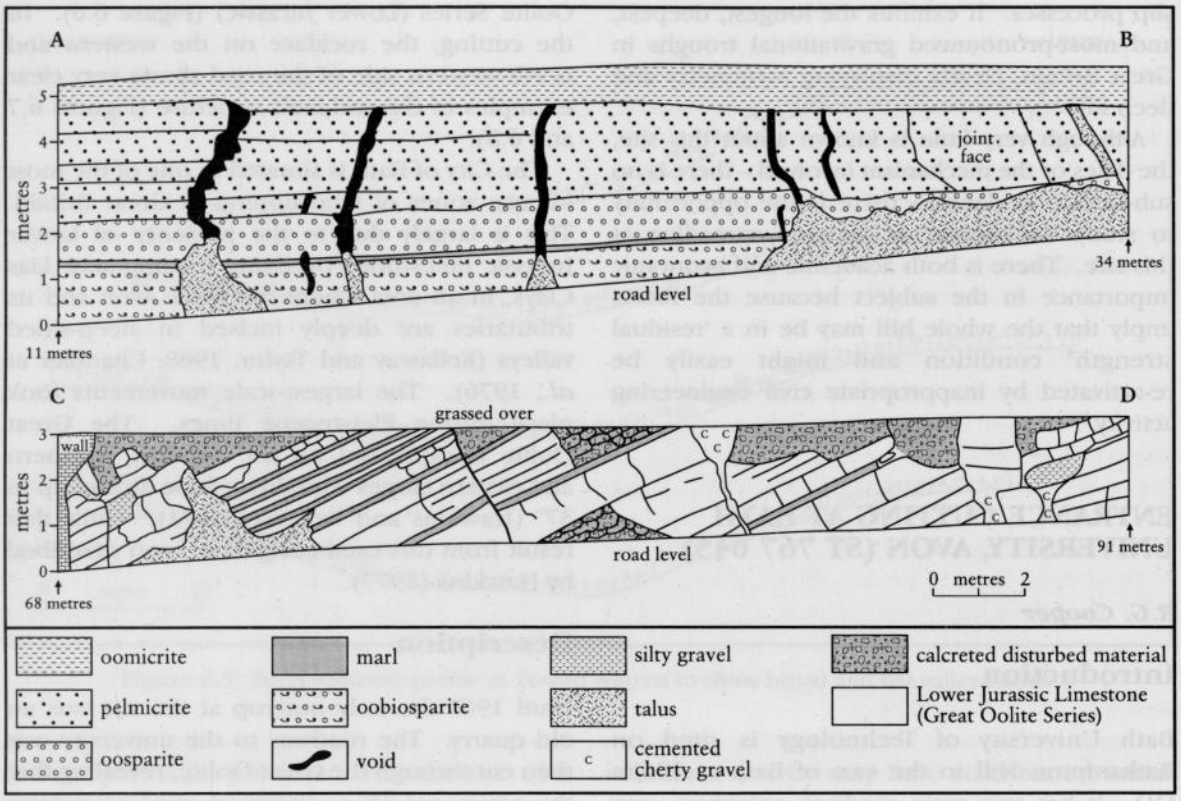


Figure 6.7 Geological sections at the Entrance Cutting at Bath University GCR site. After Hawkins (1977).

taken from the classificatory/descriptive system of Folk (1959)). Above this is another massive bed 1.9 m thick, which, especially near the road cutting, has weathered to show thickly bedded strata of oobiosparite grading upwards into oosparite. A poorly seen 0.1–0.2 m-thick band of marl in the northern cambered and collapsed section of the quarry is more clearly seen on the left of the cut entrance. In the old quarry section, this marl has frequently penetrated up into the disturbed, thinly bedded 1 m-thick overlying pelmicrite bed. Above this is a 0.4 m-thick bed of oomicrite which has a thin band of pelmicrite, and contains rounded gravel-sized clasts of micrite. In the quarry section, this intensively bored, more resistant band acts as a type of roof bed: gulls in the lower strata, often infilled with calcreted limestone fragments, frequently do not penetrate upwards through it. It is overlain by 1.25 m of cross-bedded biosparrudites, with the foresets dipping westwards. These cross-bedded rocks are overlain by a 0.35 m bed of biomicrudite, which, in the left-hand side of the cutting, is capped by 0.7 m of oosparite (Hawkins, 1977).

The structural disturbance of the beds, as revealed in the road cutting, is of considerable interest. It has not yet been possible to explain satisfactorily all of the structures, or to determine accurately the geological succession further than that given in the paragraph above. The gulls, up to 0.4 m wide on the left-hand side of the cutting 15 m from the entrance, are good examples of the way that tension applied to massive beds causes a complete fracture to open; yet in the 2 m thinly bedded upper horizon, bed-by-bed slip means that the tensional strain is taken up by many small movements and no fracture penetrates through to the surface (cf. Hawkins and Privett, 1979). The fact that these cavities exist, yet are not visible at the surface, is obviously of importance in the construction of buildings on the plateau surface (Hawkins, 1977).

Entering the cutting, one of the first things noticed is the almost horizontal bedding to the left, yet the beds on the right have a northerly dip of 26°. This face has not been sampled in detail (Hawkins, 1977). It exposes an oosparite bed, over 1 m thick with a highly bored and

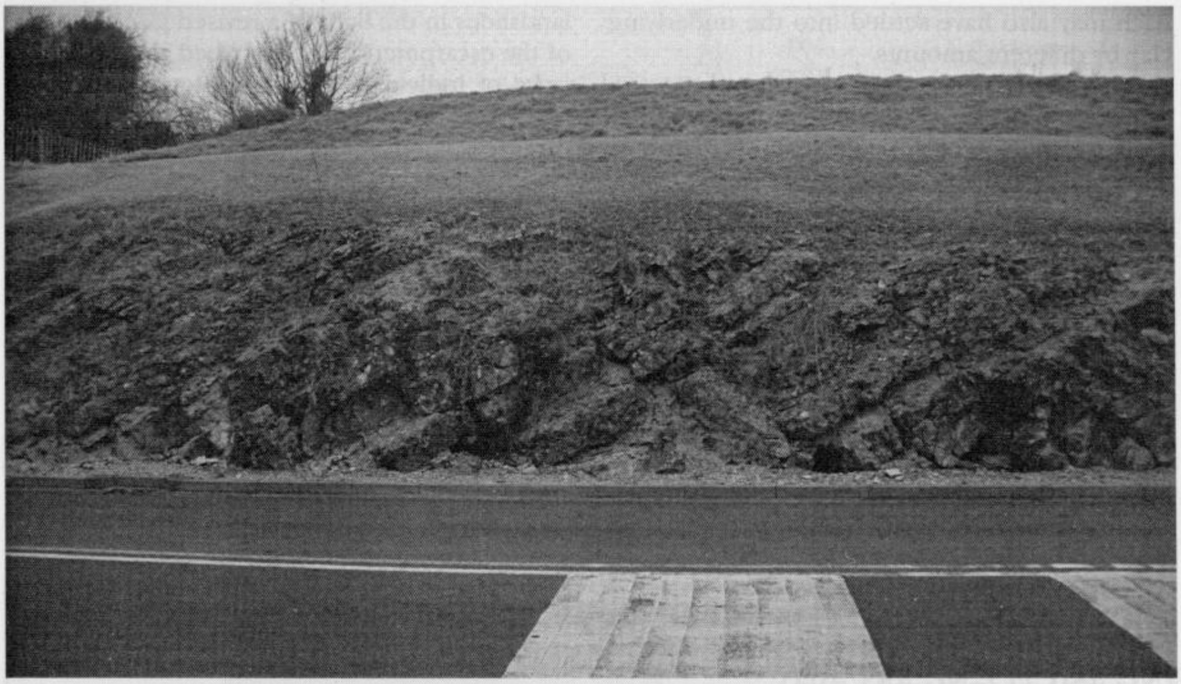


Figure 6.8 The section at the Entrance Cutting at Bath University GCR site showing the dip-and-fault structures. (Photo: R. Wright, English Nature/Natural England.)

oyster-covered surface. This distinct bed, when followed laterally is displaced in several places at the position of old gulls, now largely infilled with travertine. Ascending the incline, the sub-horizontal beds on the left-hand side suddenly begin to dip northwards, and by the bridge they have a dip of 30°. Between the sub-horizontal and inclined beds is a 0.3–0.5 m zone of disturbance, possibly representing a fault or gull breccia. Just east of the bridge, dip-and-fault structures can be seen displacing by 0.82 m and 1.7 m respectively, the 0.43 m bed of bio-sparrudite with irregular borings overlain by oosparite with fine, generally vertical, boring.

Limestones of the Great Oolite Series are seen in the old quarry at the lower end of the cutting. In the cutting section deposits of chert and flint gravel with loamy clay fill solution cavities in the limestone. Locally, one band of limestone has been dissolved out and the space infilled with gravel overlain by bedded loam. Both the limestone and the cave filling had subsequently been strongly tilted and faulted as a result of cambering and the formation of dip-and-fault structures. It therefore follows that at the time when caves were formed and the overlying drift deposits were washed into the horizontal solution cavities, the camber slope on the west side of Bathampton Down did not exist. At that time, even after the deposition of the re-sorted high-level drift deposits, this section of the Great Oolite Limestone of Bathampton Down was virtually undisturbed (Hawkins and Kellaway, 1971).

During the construction of reservoirs on the top of Bathampton Hill in 1955, some large gulls were seen, in which masses of stiff plastic clay with pebbles of patinated flint, chert and limestone, and roughly bedded loamy gravel, were enclosed in limestone fissures. In 1969 a second set of reservoirs was built on the flat plateau top. These showed large caverns, pipes and swallow holes, some having been developed by the differential solution of individual limestone bands. In these excavations the Great Oolite limestones were not cambered, the bedding of the rocks being horizontal (Hawkins and Kellaway, 1971).

Interpretation

It is clear that the gull-bounded blocks of Great Oolite have settled on the underlying Lias Clay in such a way that their downslope dips are greater than would have been expected as a result of differential lowering by the cambering process. Each individual block has rotated in the downslope direction, as well as being lowered.

Each may also have settled into the underlying clay by different amounts.

Where, as here, the downslope dip on cambered blocks exceeds the dip of the camber itself, irrespective of slope angle, the structure is termed 'dip-and-fault' (Hollingworth *et al.*, 1944). Contacts between adjacent blocks, displayed particularly well in the section at Bath, appear as normal faults heading upslope, with small downthrow on the upslope side.

Conclusions

Dip-and-fault structures can be seen at many sites in Great Britain. They are, however, displayed with particular clarity at the Entrance Cutting at Bath University GCR site.

TROTTERNISH ESCARPMENT, ISLE OF SKYE, HIGHLAND (NG 450 717–NG 481 494)

C.K. Ballantyne

Introduction

The Trotternish peninsula is the northernmost part of the Isle of Skye, the largest island in the Inner Hebrides. The peninsula is dominated by the Trotternish escarpment, a bold cliff of Paleocene lavas up to 200 m high that extends almost continuously for 23 km along almost the full length of the peninsula. Foundering of the lava scarp over the underlying Jurassic sedimentary rocks has produced the largest continuous area of landslide terrain in the British Isles, covering nearly 40 km^2 (Ordnance Survey, 1964), and two of the most spectacular post-glacial landslides in Scotland, at The Storr and Quiraing. The Trotternish landslides were first described in detail by Godard (1965), who identified an inner zone of post-glacial failures and an outer zone of subdued, hummocky slipped rock that he inferred to have been over-ridden by glacier ice. Anderson and Dunham (1966) mapped the extent of landslide terrain, all of which they regarded as post-dating the last ice-sheet, and proposed that failure had been dominated by successive deep rotational slides seated on dolerite sills. Ballantyne (1990, 1991a,b) re-evaluated the age of the Trotternish landslides in the light of a revised glacial history of the escarpment, and described the characteristics of individual landslide blocks and other landforms produced by rock slope failure along the escarpment. The age and cause of post-glacial failure at The Storr has been investigated by Ballantyne *et al.* (1998b) using cosmogenic radionuclide dating of landslide pinnacles, and the history and processes of rockfall accumulation along part of the escarpment have been reconstructed by Hinchliffe *et al.* (1998) and Hinchliffe and Ballantyne (1999) on the basis of talus stratigraphy. These investigations have thrown light on the post-glacial evolution of the Trotternish escarpment, but the deep structure and mechanics of slope failure remain more speculative.

Description

Geology and topography

During the Palaeogene Period (65–23 Ma), the opening of the North Atlantic Ocean was accompanied by prodigious volcanic activity along the western seaboard of the Scottish Highlands. This volcanicity focused between 61 Ma and 56 Ma at the end of the Paleocene Epoch. Volcanoes on Skye, Mull and elsewhere discharged large quantities of fluid, primarily basaltic lavas that buried the underlying rocks and accumulated progressively to form extensive, near-horizontal lava plateaux (Bell and Williamson, 2002; Emeleus and Bell, 2005). In the area of Trotternish, a thick succession of lava flows buried a great thickness of relatively weak sedimentary rocks (mudstones, sandstones and limestones) of Jurassic age (Bell and Harris, 1986; Hallam, 1991; Hudson and Trewin, 2002). Paleocene volcanicity was also accompanied by the intrusion of dolerite sills deep within the Jurassic strata. Jurassic sedimentary rocks and dolerite sills now underlie the low ground east of the Trotternish Escarpment, whereas west of the escarpment crest these rocks are buried under *c.* 300 m of basalts that dip gently westwards (Figure 6.9).

Anderson and Dunham (1966) have suggested that the original scarp face in Trotternish formed as a result of tilting and faulting in Neogene times. Such scarp development probably exposed the full thickness of lavas and the

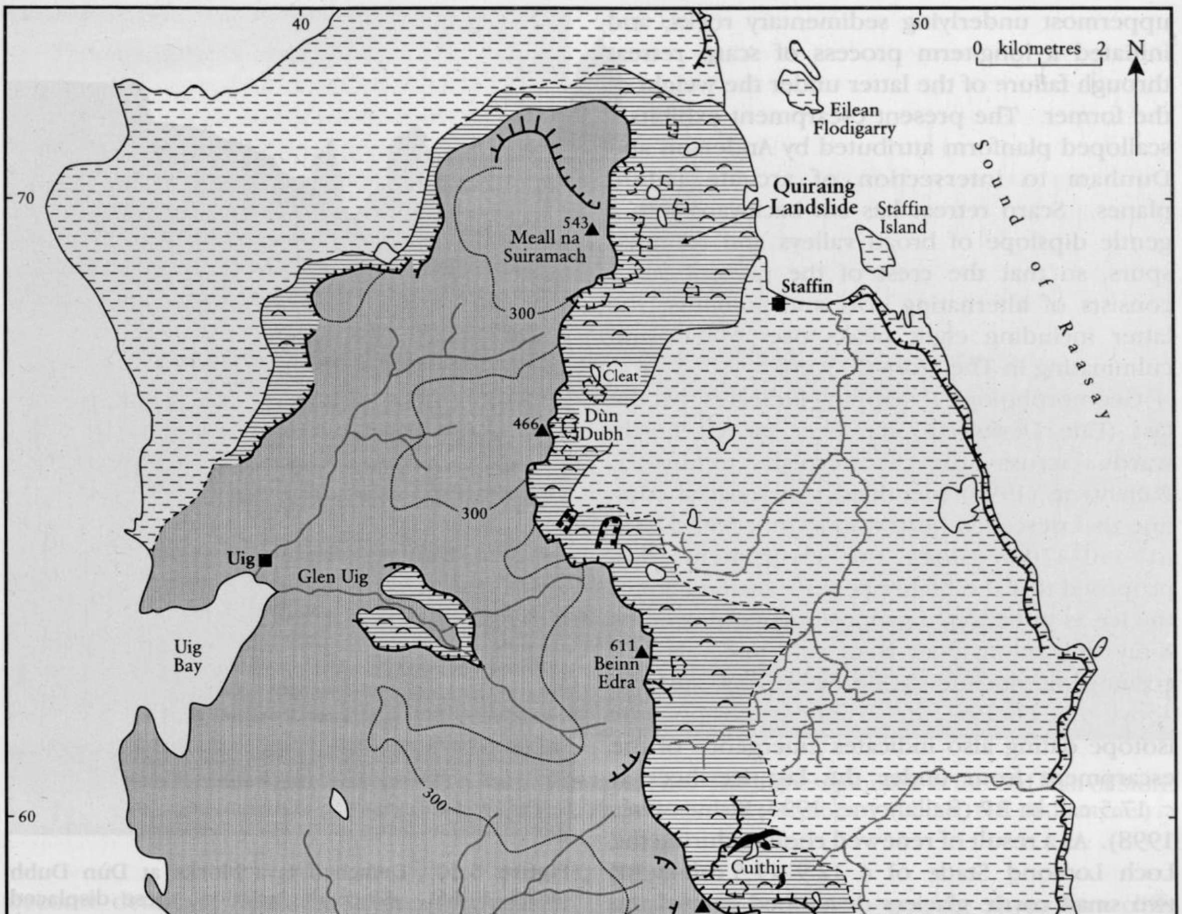


Figure 6.9 Geological and geomorphological map of the Trotternish Escarpment GCR site. The extent of landslide terrain is based partly on British Geological Survey mapping. Modified after Ordnance Survey (1964).

uppermost underlying sedimentary rocks, and initiated a long-term process of scarp retreat through failure of the latter under the weight of the former. The present escarpment exhibits a scalloped planform attributed by Anderson and Dunham to intersection of arcuate failure planes. Scarp retreat has cut backward into a gentle dipslope of broad valleys and rounded spurs, so that the crest of the present scarp consists of alternating cols and summits, the latter including eight peaks over 500 m and culminating in The Storr (719 m).

Geomorphological mapping indicates that the last (Late Devensian) ice-sheet moved northwards across the Trotternish peninsula. Ballantyne (1990) identified a periglacial trimline that descends northwards from 580–610 m to 440–470 m along the escarpment, and proposed that the higher peaks remained above the ice as nunataks, a proposition supported by X-ray diffraction analysis of clay minerals and cosmogenic nuclide dating of bedrock surfaces above and below the trimline; cosmogenic isotope dating also indicates emergence of the escarpment from under the last ice-sheet at *c.* 17.5 cal. ka BP (Ballantyne, 1994; Stone *et al.*, 1998). As a result of renewed cooling during the Loch Lomond Stade of *c.* 12.9–11.5 cal. ka BP, two small corrie glaciers developed east of the scarp, at Coire Cuithir (NG 470 592) and Coire Scamadal (NG 498 552), but the remainder of the escarpment appears to have been ice-free and exposed to severe periglacial conditions at this time (Ballantyne, 1990).

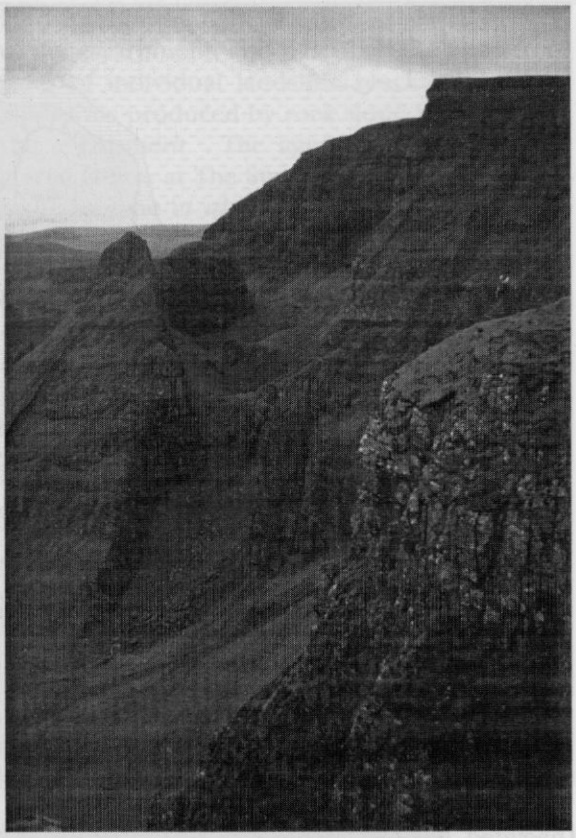


Figure 6.10 Detached lava blocks at Dùn Dubh (NG 441 666). Since deglaciation, these displaced blocks have foundered and moved laterally away from the scarp face, but without the back-tilting of lava flows characteristic of rotational sliding. (Photo: C.K. Ballantyne.)

The Trotternish landslides

Slipped rock-masses extend continuously along the foot of the escarpment over a horizontal distance of 23 km, and for a further 9 km below the basalt cliff that extends south-west from the northern end of the escarpment. A further small area of slipped rock fringes the basalt cliffs of Glen Uig on the west side of the peninsula (Figure 6.9). Two distinct zones of landsliding can be identified: an inner zone, adjacent to the scarp face, of bold angular detached blocks and pinnacles (Figures 6.10 and 6.11) and an outer zone of subdued, rounded landslip terrain, extensively covered by peat (Ordnance Survey, 1964). Anderson and Dunham (1966) suggested that erosion by the last ice-sheet removed all evidence of previous landslides, implying that all of the present area of slipped rock represents failure following ice-sheet retreat. Godard (1965), however, proposed that the outer zone represents slipped rock-masses that were subsequently over-ridden by glacier ice. Various lines of evidence favour Godard's interpretation, particularly northwards-oriented ice moulding of landslide blocks, the presence of till deposits in hollows, and over-riding of landslide blocks by lateral moraines deposited by the two corrie glaciers that developed east of the escarpment during the Loch Lomond Stade (Ballantyne, 1990). Conversely, angular blocks and delicate pinnacles of the inner zone show no evidence of glacial modification and appear to represent failure and sliding of rock masses after ice-sheet retreat.

Several different landslide forms occur within the inner zone adjacent to the basalt scarp. Adjacent to the escarpment crest are a number of incipient failures, where basalt blocks have

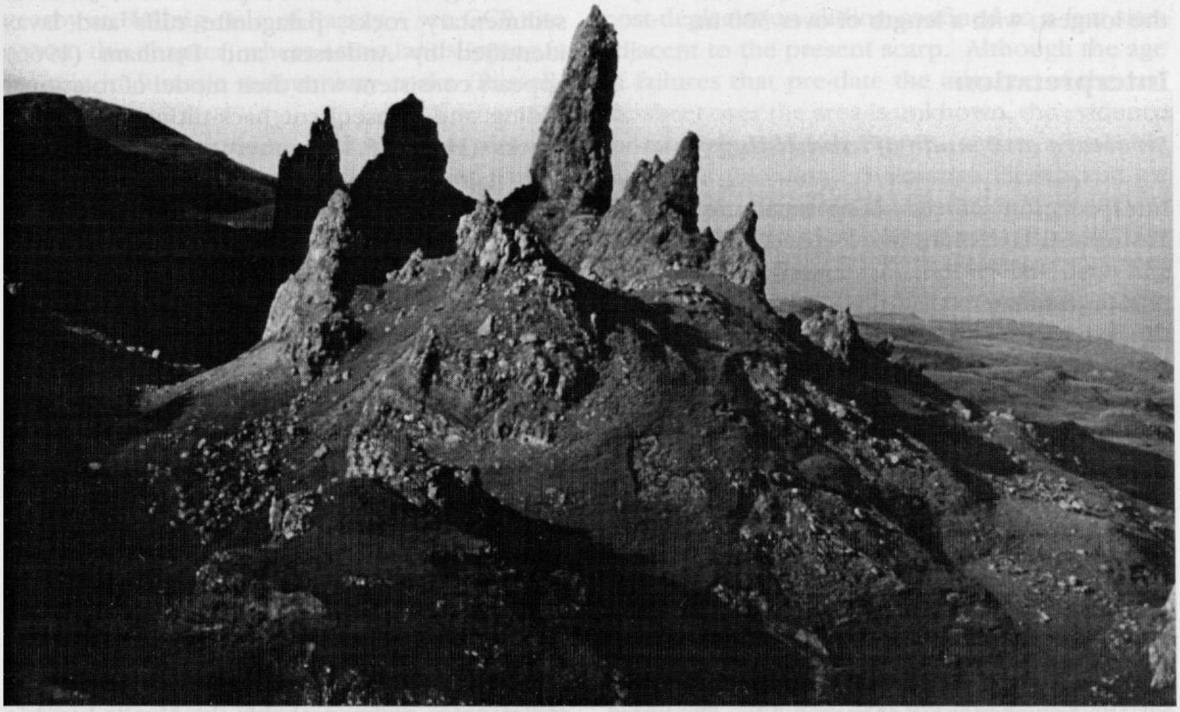


Figure 6.11 Pinnacles of shattered basalt at The Storr landslide. The highest pinnacle is the Old Man of Storr. Note the eastwards (forwards) tilt of the slipped mass, away from the escarpment. (Photo: C.K. Ballantyne.)

become detached from the crest along master joints aligned northwards, but have experienced little or no displacement. Resting against or detached a short distance from the scarp are foundered blocks of intact rock, for example at Baca Ruadh (NG 478 576), Coire Cuithir (NG 470 586) and Dùn Dubh (NG 441 666). Farther out from the crest are isolated tabular blocks such as Cleat (NG 447 669) and shattered rock pinnacles such as the Old Man of Storr (Figure 6.11) and the Quiraing needle. Along many stretches of the escarpment, however, evidence for major post-glacial rockslides is absent, and the scarp face is fringed with relict talus slopes, now extensively gullied and eroded (Hinchliffe, 1998, 1999; Hinchliffe *et al.*, 1998).

The Storr and Quiraing

The most impressive areas of landsliding occur at The Storr (NG 485 540; see also Emeleus and Gyopari, 1992) and Quiraing (NG 455 692). The entire south-east face of The Storr has collapsed to form a great hollow, Coire Faoin (NG 497 537), that is bounded to the south-west and north-west by sheer lava cliffs 200 m high. The undercliff zone of the landslide is a labyrinth of lava-capped blocks, narrow defiles and pinnacles of shattered rock, of which the 49 m-high Old Man of Storr is the largest (Figure 6.11). Below the eastern threshold of Coire Faoin, however, the landslipped blocks are subdued and rounded by glacial erosion, indicating that post-glacial landsliding was confined to an area of about 0.25 km^2 above the 350 m contour. Older, glacially modified slipped rock, however, extends down to 200 m, up to 1.5 km from the present cliff-face.

The Quiraing landslide at the northern end of the escarpment is one of the largest landslides in Great Britain. It occupies an area of *c.* 8.5 km^2 and extends 2.2 km eastwards from the scarp crest to the coastline. Like The Storr landslide, it consists of an inner (post-glacial) zone of tabular and toppled landslip blocks, pinnacles and deep clefts, and an outer zone of more subdued landslide terrain representing remnants of ancient landslides that occurred before the last and possibly earlier ice-sheets crossed the area. In the central part of the slide area, landslip blocks up to 70 m high have dammed a chain of small lakes, of which Loch Fada (NG 458 698) is the longest, with a length of over 300 m.

Interpretation

Structure and mode of landsliding

Interpretation of the deep structure of the Trotternish landslides has focused on The Storr and Quiraing slides. Anderson and Dunham (1966) estimated that prior to the failure of slipped rock at The Storr, the scarp crest lay about 600 m east of its present position. They observed repetitive outcrop of steeply dipping Jurassic sediments, palagonite tuffs and lavas in the lower parts of the slide mass, and inferred that the landslide represents successive rotational failures of a thickness of up to 300 m of sedimentary rocks and tuff under a similar thickness of basalt (Figure 6.12). They concluded that the most recent (i.e. proximal) failures were seated on an upper sill, the Creag Langall Sill, but that earler failures took place over a thicker lower sill, the Armishader Sill (Figure 6.12). The outer part of the Quiraing slide also appears to exhibit cyclic outcrop of Jurassic sediments, tuffs and lavas. Here Anderson and Dunham (1966) inferred that a thickness of *c.* 200 m of sedimentary rocks and tuff had failed under the weight of *c.* 300 m of lavas, again in the form of successive deep rotational slides, here seated on a transgressive, westward-dipping dolerite sill that crops out in nearshore islands.

Although the cyclic repetition of Jurassic sedimentary rocks, palagonite tuffs and lavas identified by Anderson and Dunham (1966) appears consistent with their model of rotational sliding and consequent back-tilting of slipped blocks (Figure 6.12), other evidence suggests that it represents over-simplification of the nature of rock displacement. It requires rotation of slipped masses against the regional westwards dip of the strata, until the outmost slipped masses are resting at improbably low angles, and infilling of the huge gaps between slipped blocks with 'rock debris' of unspecified origin. Moreover, some landslide blocks, for example at The Storr, are tilted forward (eastwards) and not backward as the rotational model suggests (Figure 6.11). The model, moreover, does not account for the detachment of large intact landslide blocks without back-tilting, notably Cleat (NG 447 669), Leac nan Fionn (NG 453 704) and Dùn Dubh (NG 459 687). These suggest a different interpretation, involving planar sliding or gliding of thick lava blocks over the weaker sedimentary rocks, probably the Upper Jurassic Staffin Shales, which in Trotternish achieve a thickness of over 117 m (Hallam, 1991). Lateral displacement and subsidence of thick lava blocks would have been accompanied by pronounced deformation of underlying sedimentary strata under the weight of the lavas; this, rather than block rotation, may account for the outcrops of steeply dipping sedimentary rocks and tuffs between the outermost lava blocks. By contrast,

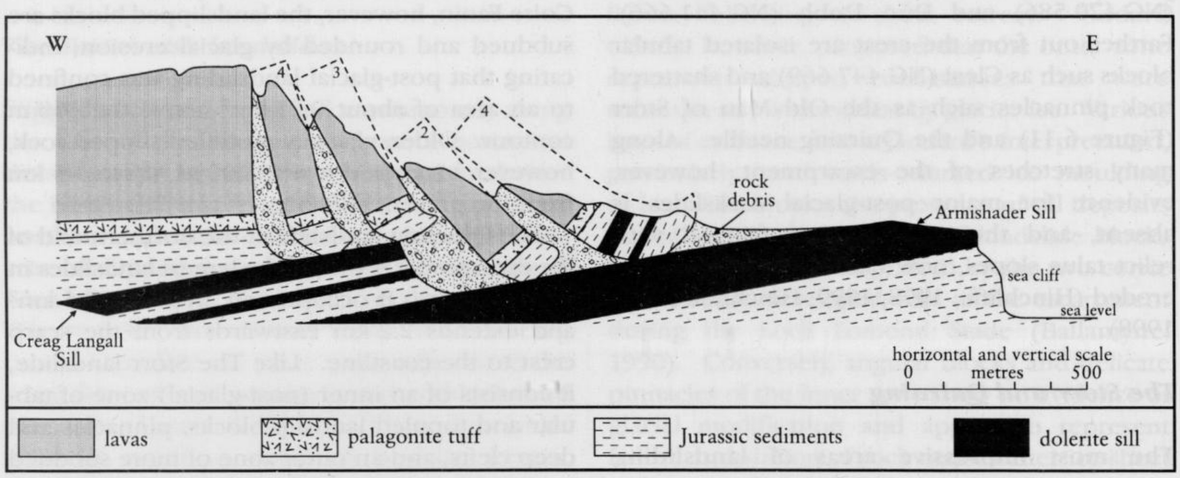


Figure 6.12 A previous interpretation of The Storr landslide, redrawn from Anderson and Dunham (1966; The geology of northern Skye, Memoir of the Geological Survey of Great Britain, p. 191, fig. 23), by permission of the Director, British Geological Survey. Their model depicts slide evolution as a sequence of successive deep rotational slides in the sedimentary rocks and palagonite tuffs underlying the basalt scarp, and involves backtitling of lava blocks. Compare with Figures 6.10 and 6.11.

backward rotation is evident in a similar context nearby at **Hallaig** (Isle of Raasay – see GCR site report, this chapter) where active landslipping is entirely in Jurassic sedimentary rocks (Russell, 1985).

The present position and altitude of such intact blocks implies displacement and subsidence without fragmentation of the lava caprock, possibly indicating gradual movement analogous to that proposed for cambering of limestone caprocks over Jurassic mudstones in England (see **Postlip Warren** GCR site report, this chapter). Like these, block movement may relate to reduction of shear strength in underlying argillaceous strata during thaw of ice-rich permafrost (Parks, 1991; Hutchinson, 1991; Ballantyne and Harris, 1994). In contrast, the shattered lava pinnacles of the post-glacial slides at The Storr and Quiraing imply catastrophic failure after deglaciation. It thus appears likely that no single mode of failure and block movement accounts for all attributes of the slipped rock-masses.

Not all rock slope failures along the Trotternish Escarpment have involved block displacement. Ballantyne (1990) mapped four smaller post-deglaciation failures, probably rock topples or translational slides, that have produced runout of coarse debris below the scarp face. South of Baca Ruadh (at NG 476 568), a tongue of vegetated debris terminates downslope in a *c.* 100 m-wide zone of bouldery mounds and hummocks, and a similar but smaller deposit occurs at NG 441 651. At NG 449 646, a jumble of limestone boulders covers an area of nearly 0.2 km^2 and records collapse of the cliff upslope. At Carn Liath (NG 464 593) a tongue of rock debris 260 m wide and 500 m long descends from 400 m to 260 m. The visible debris consists of large angular boulders, many exceeding 2 m in length. The lower part of the debris tongue extends for 250 m over a slope of only 8°–9°, and appears to be about 10 m thick. The debris tongue reflects collapse of the lava scarp upslope, probably as a small rock-avalanche, and its extended runout over gentle gradients suggests flow (debris flow or grainflow) of the landslide debris (Ballantyne, 1991b).

Timing and activity

The lateral extent of landslipped rock along the Trotternish Escarpment implies a long history of slope instability and failure, with evidence of post-deglaciation sliding confined to a few sites adjacent to the present scarp. Although the age of failures that pre-date the advance of the last ice-sheet over the area is unknown, the evidence for multiple landslide events suggests a probable link with glaciation. Successive Pleistocene ice advances northwards along the escarpment are likely not only to have steepened the scarp face and removed earlier landslide debris from the footslope, but may also have induced scarp failure due to paraglacial stress-release (Ballantyne, 2002a). It is possible that deglaciation and consequent debuttressing of the scarp face created conditions favourable for failure on several different occasions throughout the Pleistocene Epoch. Hence the broad zone of ice-moulded landslide terrain along the length of the escarpment represents a sequence of landslides that occurred in response to several episodes of deglaciation, much as the post-glacial slides at The Storr and Quiraing represent response to Late Devensian deglaciation. The long-term evolution of the escarpment may thus be envisaged in terms of alternating glacial erosional and interglacial (paraglacial) landsliding episodes.

Godard (1965) proposed that most post-glacial (inner zone) failures from the Trotternish Escarpment occurred soon after deglaciation, suggesting that they represent collapse of glacially steepened cliffs following the withdrawal of a supporting buttress of glacier ice. Cosmogenic ^{36}Cl radionuclide dating of the exposure age of two separate basalt pinnacles at The Storr landslide, however, yielded virtually identical ages averaging 6.5 ± 0.5 cal. ka BP (Ballantyne *et al.*, 1998b), consistent with radiocarbon-dated evidence for exposure of the present cliff at this time (Ballantyne, 1998). These dates indicate that the landslide occurred 7–10 ka after ice-sheet deglaciation, and at least 4 ka after the end of the Loch Lomond Stade, the final period of permafrost conditions in Scotland. Ballantyne *et al.* (1998b) inferred that progressive rock-mass weakening due to stress-release had been critical in conditioning failure, but noted that a seismic trigger could not be discounted. They also observed the significance of the age of failure in terms of indicating persistence of major landslide events well into the Holocene Epoch, an observation subsequently re-inforced by the even more recent date (*c.* 4 cal. ka BP) obtained for the **Beinn**

Alligin rock-avalanche, 30 km east of The Storr (Ballantyne and Stone, 2004). The rock avalanche at Carn Liath (NG 464 693) on the Trotternish escarpment appears to have occurred even more recently, as limited lichen cover on the source rockwall suggests failure within the past few centuries (Ballantyne, 1991b).

Although the slipped rock-masses at Trotternish appear to be stable at present, Anderson and Dunham (1966) suggested that where the toe of the Quiraing slide is being eroded by the sea there is 'continuous though not extensive movement' and that the main road near Flodigarry (NG 465 716) is 'frequently dislocated'. However, deformed sediments at the landside toe are overlain by undisturbed raised beach deposits and cut into by Holocene raised shorelines, suggesting that recent movement has been slight (Ballantyne, 1991a).

Talus accumulations

Since deglaciation, only limited areas of the Trotternish Escarpment have been affected by major rock slope failures. Much of the remainder, however, has experienced post-glacial cliff recession due to intermittent rockfall, with concomitant accumulation of talus at the scarp foot. Such talus accumulations are now essentially relict (Figure 6.13). They support an almost continuous vegetation cover and are deeply dissected by gullies, many of which exhibit evidence of recent reworking of talus sediments by debris flows. The morphology, sedimentology and stratigraphy of a section of talus near the southern end of the escarpment (NG 492 533; Figure 6.13) have been investigated by Hinchliffe (1998, 1999), Hinchliffe *et al.* (1998) and Hinchliffe and Ballantyne (1999). Calculations based on the volume of talus accumulations at the foot of the basalt cliff indicate that 4.3–7.8 m of rockwall recession has occurred since deglaciation at *c.* 17.5 cal. ka BP. The composition of the talus sediments indicates that approximately 70% of overall rockwall retreat has been due to rockfall, and that the remainder reflects granular weathering of the cliff-face. Radiocarbon-dated soil horizons within the talus imply that rockfall inputs have been very limited since the end of the Loch

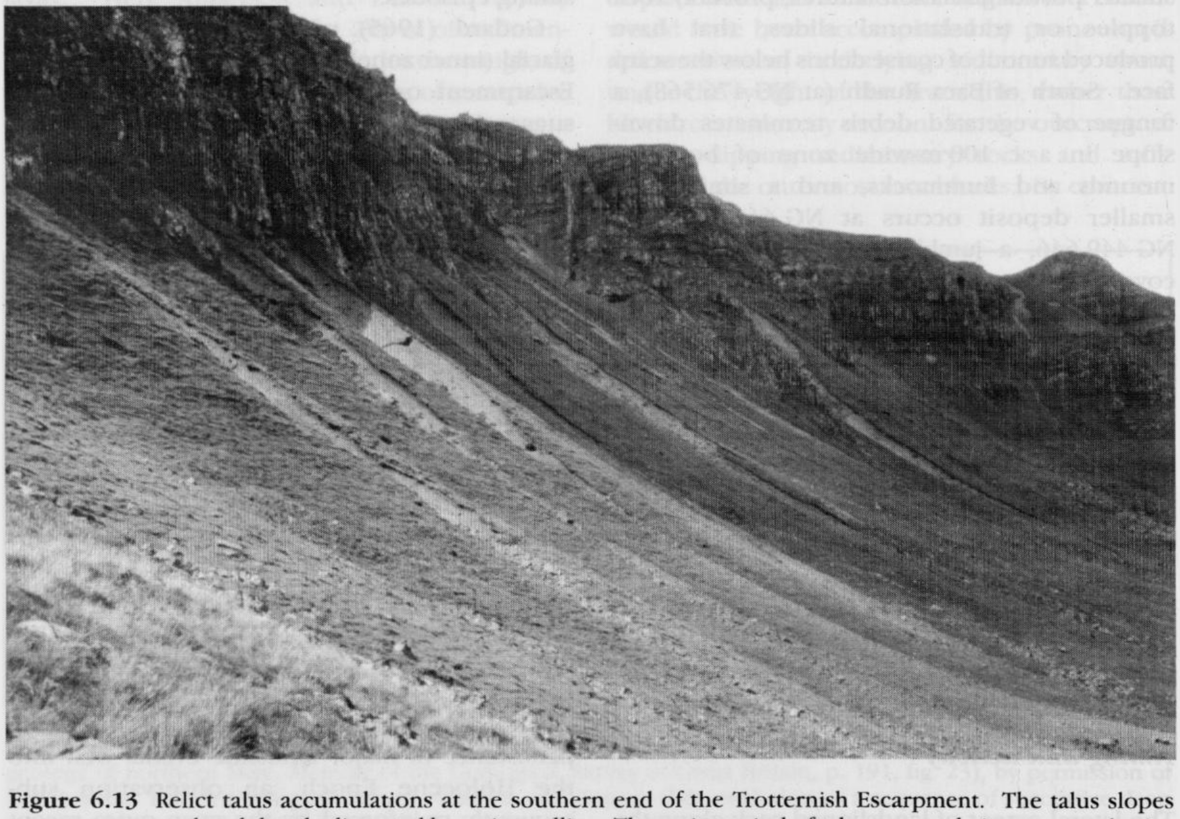


Figure 6.13 Relict talus accumulations at the southern end of the Trotternish Escarpment. The talus slopes are now vegetated and deeply dissected by active gullies. The main period of talus accumulation occurred prior to 11.5 cal. ka BP. (Photo: C.K. Ballantyne.)

Lomond Stade at *c.* 11.5 cal. ka BP, and suggest that about 80% of total rockwall retreat occurred between 17.5 cal. ka BP and 11.5 cal. ka BP, during which period rockwall retreat rate averaged about 0.75 m ka^{-1}. Hinchliffe and Ballantyne (1999) attributed this high rate of late-glacial cliff recession to stress-release following ice-sheet deglaciation, and/or frost wedging under severe periglacial conditions. The stratigraphy of the upper parts of the talus accumulations reveals stacked debris-flow horizons intercalated with occasional slopewash horizons and buried organic soils, implying that rockfall debris has been extensively reworked by intermittent debris-flows throughout much of Holocene time.

Wider significance

The assemblage of landslide features represented along the Trotternish Escarpment is characteristic of those of basaltic successions not only along the western seaboard of Scotland, but also of areas of similar geological configuration in the wider North Atlantic Igneous Superprovince (Emeleus and Bell, 2005) and elsewhere (Evans, 1984). In Scotland, less extensive but equally spectacular areas of slipped basalts overlying Mesozoic sedimentary sequences occur at several locations in the Inner Hebrides, such as Score Horan (NG 285 594) and Ben Tianavaig (NG 517 410) on Skye, and at Gribun on Mull (Bailey and Anderson, 1925; Godard, 1965; Anderson and Dunham, 1966; Richards, 1971; Ballantyne, 1986a, 1991b). In all these cases the long-term history of scarp retreat probably reflects alternating periods of glacial steepening and loading, and intervening episodes of paraglacial stress-release, rock-mass weakening, rockfall and slope failure. Many of these occur at coastal locations, raising the possibility that coastal erosion may have triggered and accelerated post-glacial failure. Some exhibit signs of recent movement (Anderson and Dunham, 1966). These landslides constitute a family of structurally and lithologically conditioned slope failures quite distinct from those found on the Precambrian or Palaeozoic rocks of the Scottish mainland (cf. 'Introduction', Chapter 2). As at Trotternish, however, the deep structure of these landslides remains uncertain.

Although much of the literature on the Trotternish Escarpment focuses on post-glacial rock slope failures and slope modification, the outer zone of older, ice-moulded landslide terrain not only provides evidence for previous episodes of slope failure during past interglacials, but also forms the most extensive area of glaciated landslide topography in Great Britain. Most pre-glacial landslides in Scotland can be identified only as modified failure scars, as the failed rock-masses and runout debris have been removed by glacier ice (Clough, 1897; Ballantyne, 2002a). Only on Trotternish have extensive areas of slipped blocks survived the passage of one or more ice-sheets, leaving a landscape of ice-moulded lava blocks and intervening lochans or peat-filled hollows. Such terrain is particularly well exemplified by the outer (eastern) parts of the Quiraing landslide.

Conclusions

The Trotternish peninsula in northern Skye includes probably the largest continuous area of landslide terrain in Great Britain (totalling nearly 40 km^2), together with the largest individual rockslide (the Quiraing slide, 8.5 km^2) and some of the finest examples of landslides associated with Paleocene volcanic rocks. These distinctions reflect the geological configuration of Trotternish, where a succession of basalt lava flows roughly 300 m thick overlies a similar thickness of Jurassic mudstones, sandstones and limestones into which have been intruded thick dolerite sills. Westward tilting and faulting of this sequence, probably in Neogene time, created a steep escarpment of lavas overlying relatively weak sedimentary rocks. Throughout the Quaternary Period, this scarp face has retreated westwards through a combination of alternating episodes of glacial erosion and interglacial landsliding and rockfall. Trotternish provides probably the clearest example in Britain of this glacial–paraglacial cycling process.

The landslide terrain east of the Trotternish Escarpment can be subdivided into two zones: an outer zone of subdued, ice-moulded landslide blocks and an inner zone of fresh tabular landslide blocks, shattered basalt pinnacles and talus accumulations. The outer zone represents landslide terrain that was over-ridden and modified by the last (and possibly earlier) ice-sheets, and represents the most extensive area of glaciated landslip terrain in Great Britain. The inner zone represents the products of scarp failure, landsliding and rockfall since ice-sheet deglaciation, which probably occurred about 17 000 years ago. Scarp failure was probably caused by steepening of the basalt cliff by glacial erosion and stress-release within the rock mass caused by unloading as the last ice-sheet downwasted. Stress-release results in slow opening of joints within the rock, resulting in progressive loss of strength. Although this effect often causes slope failure soon after deglaciation, failure may be delayed for millennia as the joint network propagates. The major landslide at The Storr occurred between 7000 and 6000 years ago, some 10 000 years after deglaciation.

Various modes of slope failure and landsliding are evident along the escarpment, but the structure of the Trotternish landslides remains speculative. Anderson and Dunham (1966) proposed that scarp retreat has been dominated by deep rotational slides in the sedimentary rocks underlying the lavas, but the configuration of some intact landslide blocks suggests planar sliding or gliding of the lava caprocks over a layer of deforming shale, possibly when the strength of the latter was reduced by thaw of ice-rich permafrost. A small number of shallow rockslides or topples have also occurred since deglaciation, and extensive areas of talus have accumulated as a result of rockfall from lava cliffs. In the southern part of the escarpment, rockfall and weathering have resulted in 4.3–7.8 m of cliff retreat since deglaciation, most of which occurred in the interval between ice-sheet deglaciation and the end of severe periglacial conditions around 11 500 years ago. The resulting talus slopes were subsequently modified by intermittent debris-flows, and are now essentially relict and extensively eroded.

Much greater amounts of post-deglaciation cliff retreat have occurred at major landslide sites, notably at The Storr and Quiraing. Both of these famous landslides consist of a chaotic inner zone of tilted lava blocks, deep defiles and shattered lava pinnacles, and an outer zone of ancient landslide blocks modified by the passage of the last and probably earlier ice-sheets. The Storr landslide extends 1.5 km from scarp crest to toe, and Quiraing landslide reaches the sea 2.2 km from the escarpment crest. The toe of the latter is reported to have experienced recent movement, possibly due to coastal erosion.

HALLAIG, ISLE OF RAASAY, HIGHLAND (NG 588 387)

R.G. Cooper

Introduction

The Isle of Raasay lies between the Isle of Skye and the Applecross peninsula of the Scottish mainland. On its eastern side, at Hallaig, is a large landslip (Figure 6.14), which has moved in recent times (a slip was recorded in 1934). It is

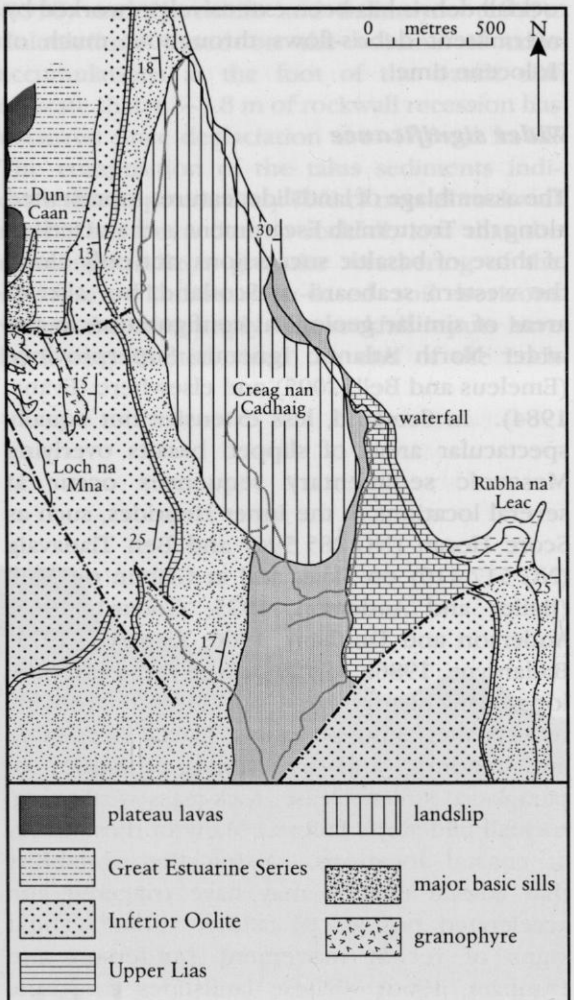


Figure 6.14 The geological setting of the Hallaig landslide on the coast of the Isle of Raasay.

about 1.8 km long from north to south, and extends from the Cadha Carnach cliff on the east side of Dun Caan (NG 583 384) for 800 m to the coast. The landslip has been mapped in detail by Russell (1985, see Figure 6.15), and much of his work is utilized in this account.

Description

The Hallaig landslip lies on the eastern side of Raasay, immediately south-east of Dun Caan (NG 579 395), adjacent to the Inner Sound. The landslip forms a large crescent-shaped topographical feature extending from 290 m above OD to sea level. The main backslope to the slip is itself 150 m high in places (e.g. Cadha Carnach) and a major bench at *c.* 150 m above OD represents the top of the slipped mass. Loch a' Chada-charnaich infills a hollow on this bench. The landslipped material forms a steep slope from *c.* 150 m above OD to sea level,

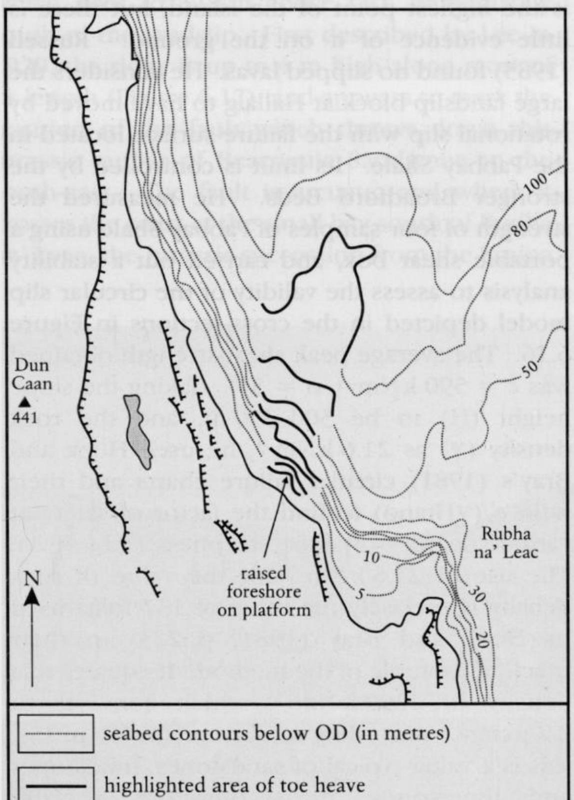


Figure 6.15 Slipped masses of the Hallaig landslips and the offshore features. After Russell (1985).

constituting a very large mass of failed material. The crags of Creag nan Cadhaig are prominent on this lower slope. They lie immediately to the north of the deserted village of Hallaig and are formed of Jurassic limestone, containing fissures (probably widened joints) of unknown depth, which are partly open to the surface. The fissures are up to 4 m wide at the surface, and appear to have opened in response to internal deformation of the slipped rock mass.

The most significant aspect of Russell's (1985) mapping of the landslide was the recognition, in the field, of the stratigraphical sequence on the backface of the landslip, and its repetition in the slipped mass. The critical sequence is:

Bearreraig Sandstone Formation
Raasay Ironstone Formation
Scalpa Sandstone Formation

In the field he mapped this sequence both along the backface and again across the landslip. Allied with measurements of dip, this provided data for five cross-sections (Figure 6.16). In the backface, he found the in-situ bedding dipping between 12° and 15° west, while in the slipped mass the bedding strikes north–south with dips between 44° and 58° west.

In the vicinity of the sea cliffs he found that the Pabbay Shale, which underlies the Scalpa Sandstone, has exactly the changes in dip shown in Figure 6.16. One small exposure of Pabbay Shale exhibits sheared surfaces and slickensided surfaces, the latter cutting one another. In the same area the underlying Broadford Beds have the normal tectonic dip, and hence were not involved in the slip at this location. Russell (1985) pointed out that near the foreshore the Pabbay Shale is raised, indicating an area of toe heave. The Admiralty Chart of the area (Chart No. 2480) shows tongue-like sea-floor features suggesting that this area of toe heave continues offshore.

Russell (1985) observed that open joints are a distinctive feature of the Raasay landslides, and accordingly made a study of them. In the area of the landslide itself, they are to be found only in the Bearreraig Sandstone Formation (here, a series of coarse-grained calcareous and non-calcareous sandstones). The joints generally have an east–west trend, with a few in the south of the landslide block trending north–south.

To the west of the southern part of the landslip block, widened joints in the Scalpa Sandstone have grassy bottoms and show no

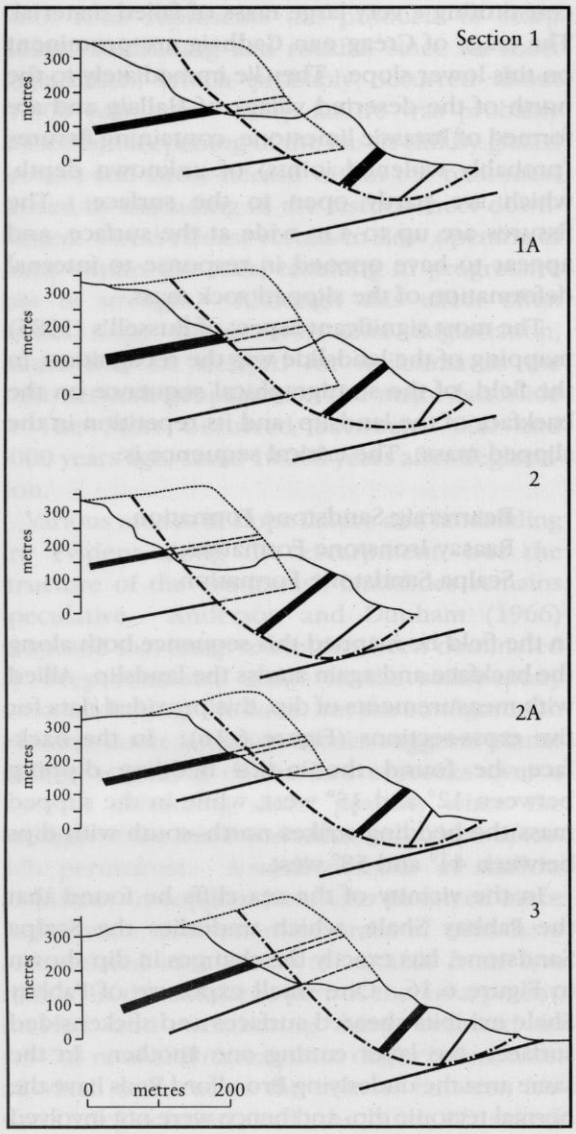


Figure 6.16 Cross-sections of the Hallaig landslip. After Russell (1985).

displacement across bedding units. On average they are 6.65 m wide, which Russell (1985) attributes in large part to subaerial erosion of their walls. On average they are 1.5 times as deep as they are wide. Their average weighted trend is 343°, approximately parallel to the slope trend. Many of the widened joints in the Bearreraig Sandstone Formation in the same area are narrower and deeper than those in the Scalpa Sandstone, being on average 5.4 times as deep as they are wide. Their average weighted trend is 349°, again running approximately parallel to the slope. Those close to the crest of the landslide's backface show some downward vertical displacement of their eastern walls (Russell, 1985, p. 59).

Interpretation

Despite the similarity of its east-facing position, the landslip at Hallaig on the Isle of Raasay differs greatly in its geological setting from the **Trotternish Escarpment** landslips on the Isle of Skye. Anderson and Dunham (1966) suggest that within the 145 m pile of sedimentary rocks exposed in the backface of the landslip, the first movement was that of the Portree Shale and Raasay Ironstone formations strata overlying the competent Scalpa Sandstone. This was followed by slipping of the Pabbay Shale with the Scalpa Sandstone acting as over-burden, and then by failure of the Great Estuarine Series and overlying Paleocene lavas.

Such a sequence of events may be necessary to explain the development of the entire slope eastwards from the summit of Dun Caan, which is the highest point of the island, but there is little evidence of it on the ground. Russell (1985) found no slipped lavas. He considers the large landslip block at Hallaig to have moved by rotational slip with the failure surface located in the Pabbay Shale. Its limit is controlled by the stronger Broadford Beds. He measured the strength of four samples of Pabbay Shale using a portable shear box, and carried out a stability analysis to assess the validity of the circular slip model depicted in the cross-sections in Figure 6.16. The average peak shear-strength obtained was $c = 590\ \mathrm{kNm^{-3}}$, $\phi = 35°$. Taking the slope height (H) to be 300–350 m, and the rock density (γ) as $21.6\ \mathrm{kNm^{-3}}$, he used Hoek and Bray's (1981) circular failure charts and their ratio $c'/(\gamma H \tan\phi)$ to find the factor of safety at various degrees of slope steepness (Table 6.1). The use of $21.6\ \mathrm{kNm^{-3}}$ for the value of rock density is precisely the value of 137 $\mathrm{lb/ft^3}$ used by Hoek and Bray (1981, p. 223) in their practical example of the method. It equates to a density of $2200\ \mathrm{kNm^{-3}}$, which converts to $2.2\ \mathrm{g\,cm^{-3}}$. According to Farmer (1968, p. 15), this is a value typical of sandstones, mudstones and limestones; it is therefore a valid approximation to use. Russell (1985) concluded that failure could occur at an angle of 64° with a slope height of 300 m, and at an angle of 58° with a slope height of 350 m. This is in good agreement with slope gradient reconstructions shown on the cross-sections (Figure 6.16).

Table 6.1 Factors of Safety (*F*) for the Hallaig landslip. After Russell (1985).

Slope height (m)	*Slope angle (degrees)*	*F*
300	80	0.75
300	70	0.90
300	60	1.06
350	80	0.69
350	70	0.84
350	60	0.97
350	50	1.12

The Hallaig landslip also differs in its geological setting from the **Trotternish Escarpment** landslips in that features in the immediate vicinity suggest that recent tectonic movements have taken place on the east side of Raasay. Such movements are likely to have had some effect on the landslip. The most apparent sign of such movement is a regular ridge of sandstone blocks running along the north-western side of Beinn na' Leac, a flat-topped hill of Bearreraig Sandstone (Aalenian and Bajocian, Mid-Jurassic, age) which rises to 319 m above OD, immediately south of the landslip. First described by Lee in 1920, the ridge is up to 4 m high along most of its length (Figure 6.17), and appears to mark the position of the fault which throws down this Jurassic outlier of Bearreraig Sandstone to the south-east. The fault is arcuate, and where it crosses the coast at the small bay south of Rudha na' Leac, the Jurassic succession from the Scalpa Sandstone (Upper Pliensbachian) to the Bearreraig Sandstone to the south is juxtaposed against the Broadford Beds (Sinemurian and Hettangian) to the north; the throw of the fault is of the order of 300 m (Morton, 1969). Lee (1920) comments that the materials forming the ridge must be derived from the crags immediately upslope, and interprets the feature as scree accumulations now separated from the slope by movement on the fault. Since the ridge is too fragile to have survived glaciation, this movement must have been post-glacial, and the ridge is so fresh in appearance that the movement appears to have occurred recently. It may be a late expression of isostatic re-adjustment to the removal of the Devensian glaciers. Further evidence of recent movement is provided by features of Beinn na' Leac itself, and by reports quoted by Anderson and Dunham (1966). The east side of Beinn na' Leac is made up of

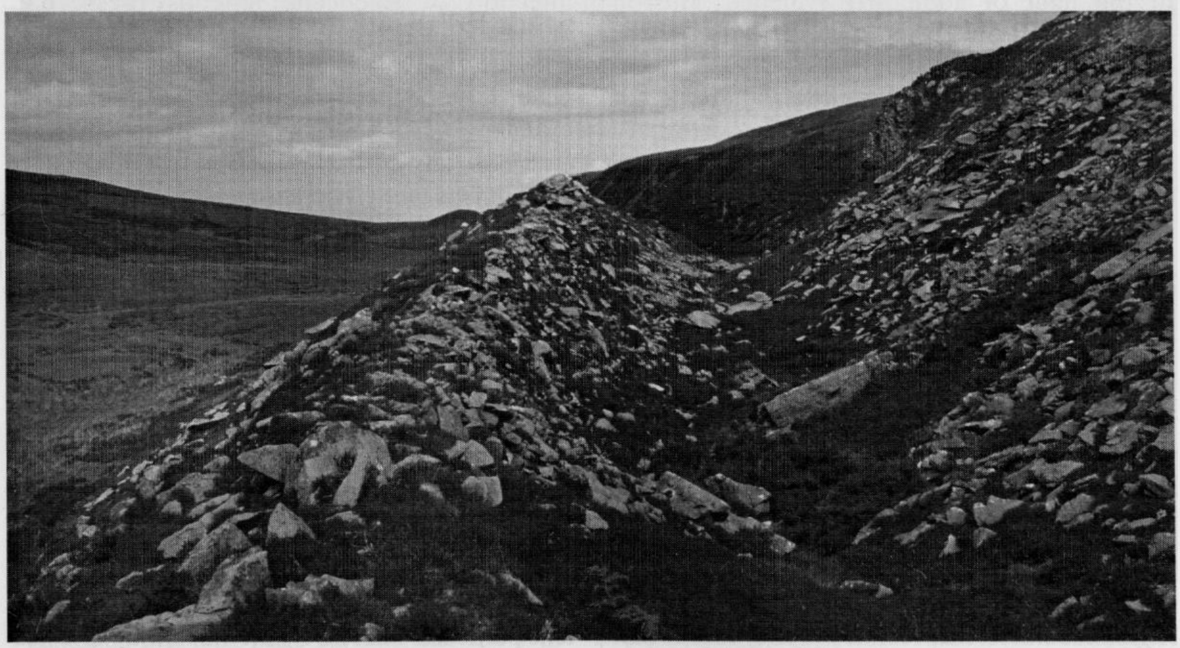


Figure 6.17 The Hallaig landslide – Beinn na' Leac ridge. (Photo: R.G. Cooper.)

landslips, to the extent that they greatly increase the apparent thickness of the Scalpa Sandstone that crops out there (Lee, 1920). Many vertical fissures of great depth occur on the surface of Beinn na' Leac. The latter are wide enough to be entered, and have been partly explored by cavers.

Evidence of recent movement, as adduced by Anderson and Dunham (1966), includes the local newspapers for 7 August 1934, which reported that twice within a period of six weeks a volcanic eruption had taken place in Raasay. Rising steam and smoke, showers of stones and a loud rumbling noise were reported by local inhabitants. They also reported that Professor A.D. Peacock, who visited Raasay shortly after the most recently reported occurrence, attributed these phenomena to stones falling down one of the very extensive fissures backing the latest slipped mass. Anderson and Dunham (1966) comment that it is 'more probable that such a spectacular disturbance was due to movement on a larger scale, i.e. to renewed slipping of the unstable mass'. It can be added that renewed movement on the fault could also give rise to such phenomena, and indeed to renewed slipping of the unstable mass.

It may be worthy of note that Musson *et al.* (1984) have traced evidence of an earthquake that took place on the nearby mainland on 16th August 1934, nine days after the newspaper reports noted by Anderson and Dunham (1966). They placed its epicentre in Strathconon Forest. In the light of additional evidence, Musson (1989) placed the epicentre farther west, in the Torridon area, although the position is still poorly determined. The earthquake was felt over a wide area, and Neilson and Burton (1985) list it as one of the larger British earthquakes of the 20th century, with an instrumental magnitude of 3.8 ML, which Musson (1989) regards as a rather small estimate considering the large area over which it was felt. Torridon village, assumed by Musson (1989) to be close to the epicentre, is 35 km from the Hallaig landslip on the Isle of Raasay. Anderson and Dunham (1966) note that in the mid-1950s many new fissures could be seen, as could evidence of 'considerable and recent movement', and Russell (1985) provides photographs of manifestly recent shallow slides (his plates 8 and 9). In this connection it is relevant to record that Ballantyne (1997) in a review of rock slope failures in the Scottish Highlands noted that Holmes (1984) found that most translational rockslides had taken place over failure planes inclined below the residual angle of friction, the lower threshold angle for rock masses to slide under their own weight. After considering and eliminating glacial over-steepening, progressive failure, and high cleft-water pressures as causes of rock slope failures under these conditions, Ballantyne (1997) concluded that seismic activity was a likely triggering factor.

One further circumstance may be significant in relation to the Hallaig landslip. This is the remarkable depth of the sea-channels which flank Raasay. In particular, the Inner Sound, which lies between Raasay and the Applecross peninsula of mainland Scotland, has a channel which includes the deepest submarine hollows in the British sector of the continental shelf (Whittow, 1977). These extend down to more than 300 m below sea-level. The submarine slope is steep, with an average angle of about 19°, although the eastern side of the trench is seldom steeper than about 8° (Robinson, 1949; confirmed by the bathymetric survey of Chesher *et al.*, 1983). Thus the steeper side of the trench lies adjacent to the east coast of Raasay, and the Hallaig landslip. The asymmetry of the trench led Sissons (1967) to write: 'a fault control of this trench seems likely and it may well be that the fault has caused these relatively soft rocks [i.e. the Jurassic sedimentary rocks] to form the floor of part of the Inner Sound and so aid its excavation [to these great depths] by glacier ice', even though they have been strengthened by the injection of Paleocene igneous rocks, e.g. dolerite sills and dykes. Whittow (1977, p. 276) put forward the opinion that the presence of major faults on Raasay suggests that the entire chain of islands (Scalpay, Raasay, Rona) may be fault controlled: 'Their entire eastern shores may have been carved from faultline scarps, and the neighbouring ocean deeps excavated along the relatively soft rocks of the down-faulted [Jurassic] sedimentary basins'.

Conclusions

The Hallaig landslip on Raasay is a very large crescentic failure of Jurassic sedimentary rocks that shows evidence of a classic circular failure. Unlike the **Trotternish Escarpment** landslips of northern Skye, Paleocene lavas were not of importance here, at least in the first two of the landsliding stages postulated. The landslip is lent particular interest by the manifestly unusual events (for Great Britain) apparently taking place at Beinn na' Leac, immediately to the south. Here, inferred fault movement, detachment of scree, and widening of joints all seem to point to some kind of post-glacial flexure and faulting of the sandstones which make up Beinn na' Leac itself. These tectonic events may be related to the great depth of the Inner Sound, enhanced by glacial excavation. These factors and the various landslips of the area seem intimately related, in ways still to be evaluated in detail. However, the Hallaig landslip is the only British mass-movement site with good evidence for neotectonic activity.

Shocks accompanying submarine slumping on the 19° underwater slope might have been responsible for the phenomena observed in Raasay in 1934, and could also have acted as an episodic trigger of continued movements of the Hallaig landslip. Both subaerial movement of the landslip, and submarine slumping, could be triggered by fault movement, and it is known that the general area was seismically active in 1934.

AXMOUTH–LYME REGIS, DEVON–DORSET (SY 257 897–SY 333 915)

R.G. Cooper

Introduction

The Axmouth–Lyme Regis stretch of the south coast (Figure 6.18) comprises one of the best known areas of landslipping in Great Britain: it includes the site of arguably the first large-scale landslide ever to have been the subject of detailed scientific description by geologists, and it was the mass-movement site most widely suggested for inclusion in the Geological Conservation Review (Cooper, 1982). It is a National Nature Reserve (declared in 1955), selected primarily for its geological interest, especially its landslides. There has been considerable debate about the mechanisms responsible for the landslides and the develop-

Figure 6.18 The Axmouth to Lyme Regis Undercliffs region. This photograph shows the famous Bindon Landslide that took place on Christmas Eve 1839. It is probable that this is the first landslide to be fully described in a scientific memoir. (Photo: courtesy of http://www.ukaerialphotography.co.uk.)

ment of the complex of landforms found there at the present day.

The site is about 9.5 km in length from west to east, and generally extends inland from the high-water mark for about 500 m. The strata involved consist of a series of easterly dipping early Mesozoic argillites and 'limestones' (Keuper and Rhaetic (Upper Triassic) and Lower Lias (Lower Jurassic)), successively exposed by faulting. These are overlain unconformably by arenaceous and calcareous sediments (Gault and Upper Greensand (Lower Cretaceous) and Chalk (Upper Cretaceous)). The plane of unconformity dips just east of south at about 5°. The present in-situ sea cliffs on the coastal boundary of the area are formed of a more limited range of strata: Keuper Marls, the White Lias division of the Rhaetic, and the Blue Lias division of the Lower Lias (Figure 6.19; Pitts, 1979).

Description

(a) General

There is a continuous series of landslips along the coast from Axmouth to Lyme Regis, named successively Haven Cliff, Culverhole Cliffs, Bindon Cliffs, Dowlands, Rousdon Cliff, Charton Bay, Whitlands, Pinhay Bay and Ware Cliffs (Figure 6.20). All have histories of large-scale landslipping throughout post-glacial times, and have displayed similar features (Pitts, 1982, 1983a). Although the major component in most or many of the slips has probably been rotational, detailed examination by Pitts (1979) has shown that a wide variety of mass-movement types are present. These include:

1. Rockfalls caused by undermining of relatively competent rocks by erosion of relatively incompetent horizons, typified by falls of Blue Lias calcarenites in the sea cliffs at Pinhay Bay.
2. Rockfalls and clayfalls caused by frost, water or desiccation in multi-jointed or fissured materials. The scale of the falls varies with the frequency of discontinuities, between the relatively widely spaced major fractures of the Chalk and the indurated Upper Greensand facies in the cliffs in the back of the undercliff, to the closely fissured Keuper Marls of Haven Cliff and Culverhole Cliffs.
3. Gully enlargement associated with cliff-top seepage points, as in the Keuper Marls of Culverhole Cliffs.
4. Forward toppling of columns of rock bounded by approximately vertical, continuous fractures, on the seaward edge of Goat Island.
5. Successive rotational slips, as in the Chalk–Upper Greensand–Lower Lias succession of Pinhay Bay, or the Chalk–Upper Greensand–Lias–Rhaetic succession of Charton Bay. These are all renewed movements of the original slipped masses.
6. Retrogressive slips, as in the Chalk–Upper Greensand succession in The Chasm at Bindon Cliffs.
7. Non-circular to translational slides, as at Bindon Cliffs, leaving a relatively undisturbed slipped mass. These represent re-activation of slipped masses.
8. Debris slides with weathering or depositional discontinuities and a mainly

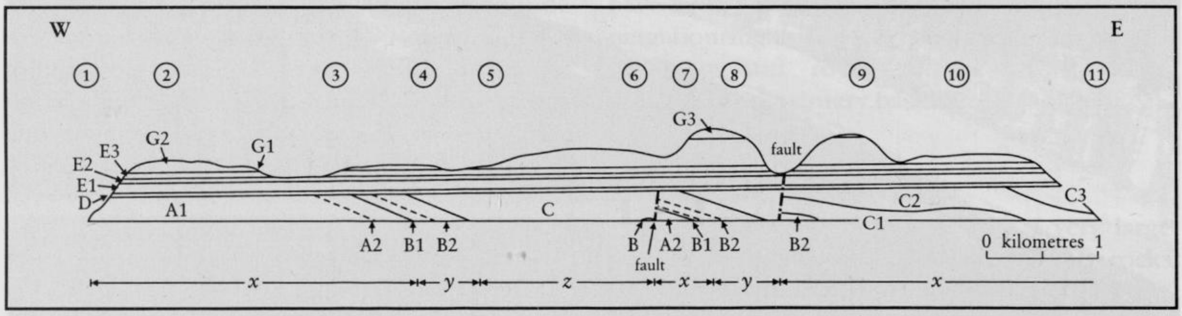


Figure 6.19 Schematic geological section of the coast between Axmouth and Lyme Regis. (1) River Axe; (2) Haven Cliff; (3) Culverhole Cliffs; (4) Bindon Cliffs; (5) Dowlands; (6) Rousdon Cliff; (7) Charton Bay; (8) Humble Point; (9) Pinhay Bay; (10) Ware Cliffs; (11) Lyme Regis. Geological succession: G3 – Upper Chalk; G2 – Middle Chalk; G1 – Cenomanian limestone; E3 – Phosphatic Upper Greensand; E2 – Cherty Upper Greensand; E1– Foxmould; D – Gault; C2 – Shales-with-Beef; C1 – Blue Lias; B2 – Lilstock Formation; B1 – Westbury Formation; A2 – Blue Anchor Formation; A1 – Red and Variegated Marls of the Mercia Mudstones Group. After Pitts (1979).

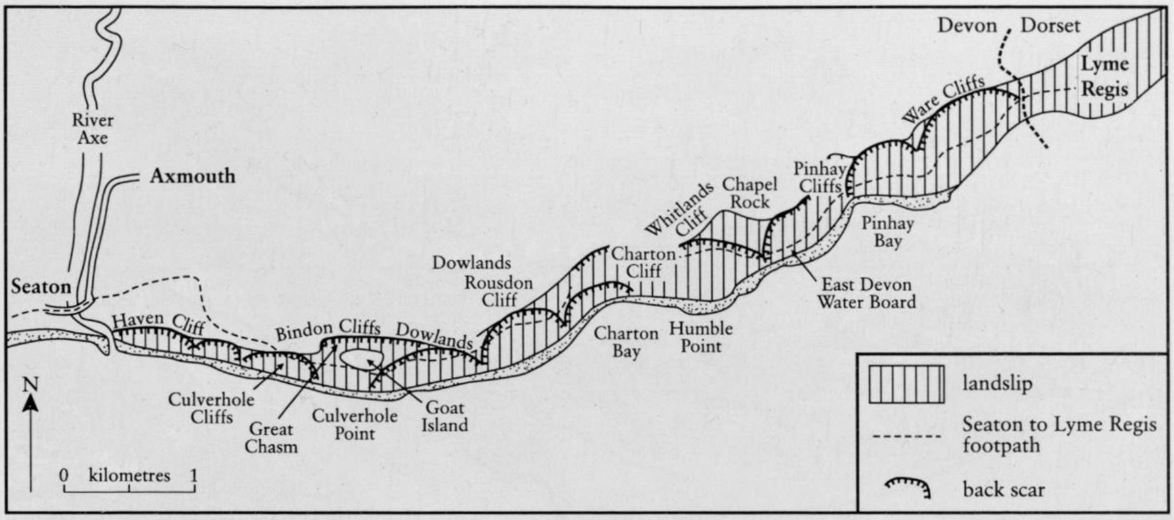


Figure 6.20 The landslides of the Axmouth to Lyme Regis Undercliffs National Nature Reserve. After Pitts and Brunsden (1987).

disturbed slipped mass, as at the toe failures of Haven Cliff, and the scree-slope failures in front of slipped Chalk and Upper Greensand blocks at Charton Bay.

9. Mudslides: these are mainly toe features, being the terminal stage of successive rotational slipping where adequate comminution of debris, clay bedrock and seepage tend to occur together, as at the cliff-top at Pinhay Bay, and at beach level at Dowlands.
10. Radial heave and slow, creep deformation at the toes of mudslides at both Dowlands and Pinhay; and small cliff-foot flows from saturated talus beneath seepage points, as at Haven Cliff and Ware Cliffs.
11. Liquefaction: structural collapse of Foxmould and subsidence of overlying strata, as at depth beneath Goat Island in the Bindon slip.
12. Sand-runs: collapse of dried-out non-cohesive arenaceous deposits, especially Foxmould, as at Charton Bay.

At Charton Bay the landform is unusual: the undercliff is two-tiered, i.e. there have been two separate landslips, one above the other, within the 127 m-high cliffs (Pitts, 1986). The upper undercliff is probably of great antiquity, while the lower undercliff formed as recently as 1969.

Two large-scale slips are very well-documented; the slip at Bindon in 1839, and the slip at Whitlands in 1840.

(b) The 1839 slip at Bindon

The Bindon area has a long and very complex history of landslipping, which has been pieced together in great detail by Pitts (1982, 1983a), using documentary sources. Probably the best-documented event, because it has been the most spectacular to take place in modern times, was the landslip at Dowlands and Bindon in 1839 (Pitts, 1974, 1982). Numerous eyewitness accounts of this slip have been documented (e.g. Roberts, 1840), and many illustrations are still extant (Figures 6.21–6.24).

The slip is particularly remarkable for its cross-sectional shape. The new main cliff-face at the rear (landward) side was then up to 64 m high, and has in front of it a large depression ('The Chasm') into which about 8 ha of land had subsided. The length of The Chasm is about 800 m, while its breadth increases from 60 m in the east to 120 m in the west. The amount of foundered material, of which some is back-tilted, was estimated at 4.2×10^6 m^3, weighing nearly 8.1×10^9 kg. Most of this is broken into a jumble of small rifted masses and pinnacles of rock, but in places blocks of about one hectare remain intact though tilted. Beyond The Chasm a counterscarp of Chalk borders an isolated upstanding area of 6 ha, which soon became known as 'Goat Island'; it had moved seawards and subsided to some extent, but corn- and turnip-fields and hedges survived (Hutchinson, 1840) (Figure 6.25).

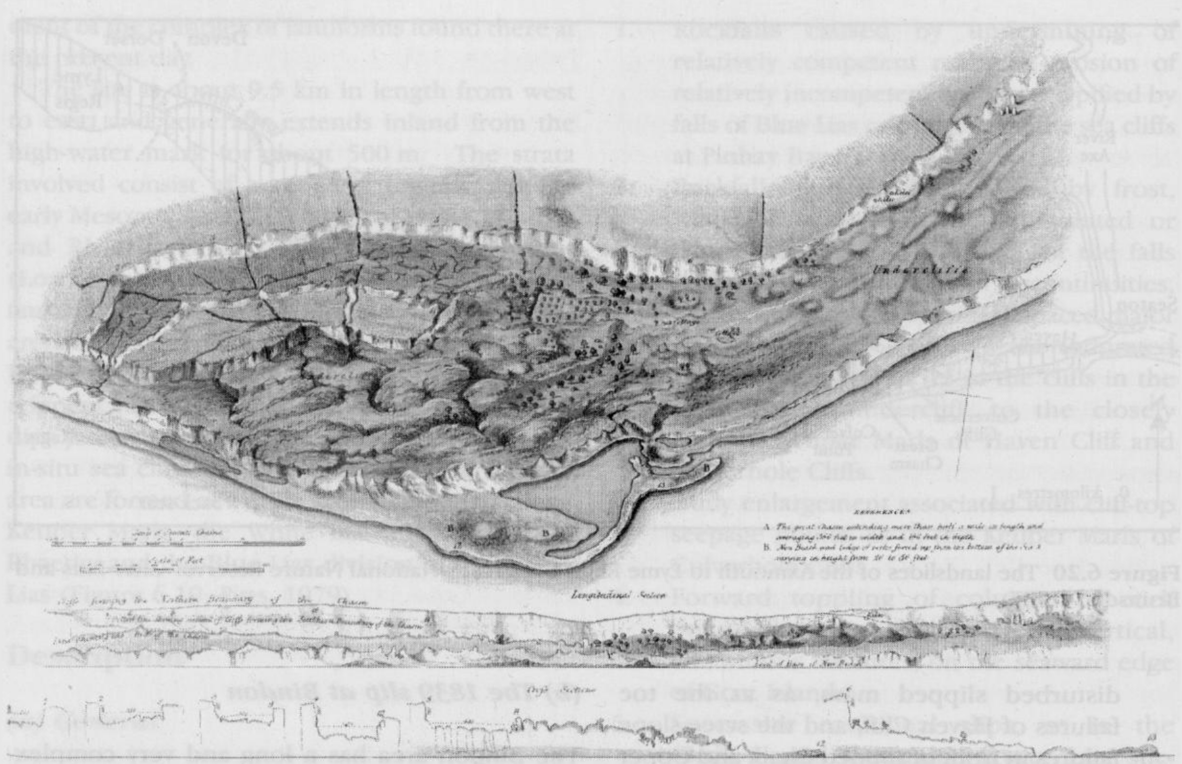


Figure 6.21 Ground plan and section of the Bindon Landslip (1839). From Conybeare *et al.* (1840), reproduced with permission of Lyme Regis Museum.

Figure 6.22 A view of 'The Chasm' looking west. From Conybeare *et al.* (1840), reproduced with permission of Lyme Regis Museum.

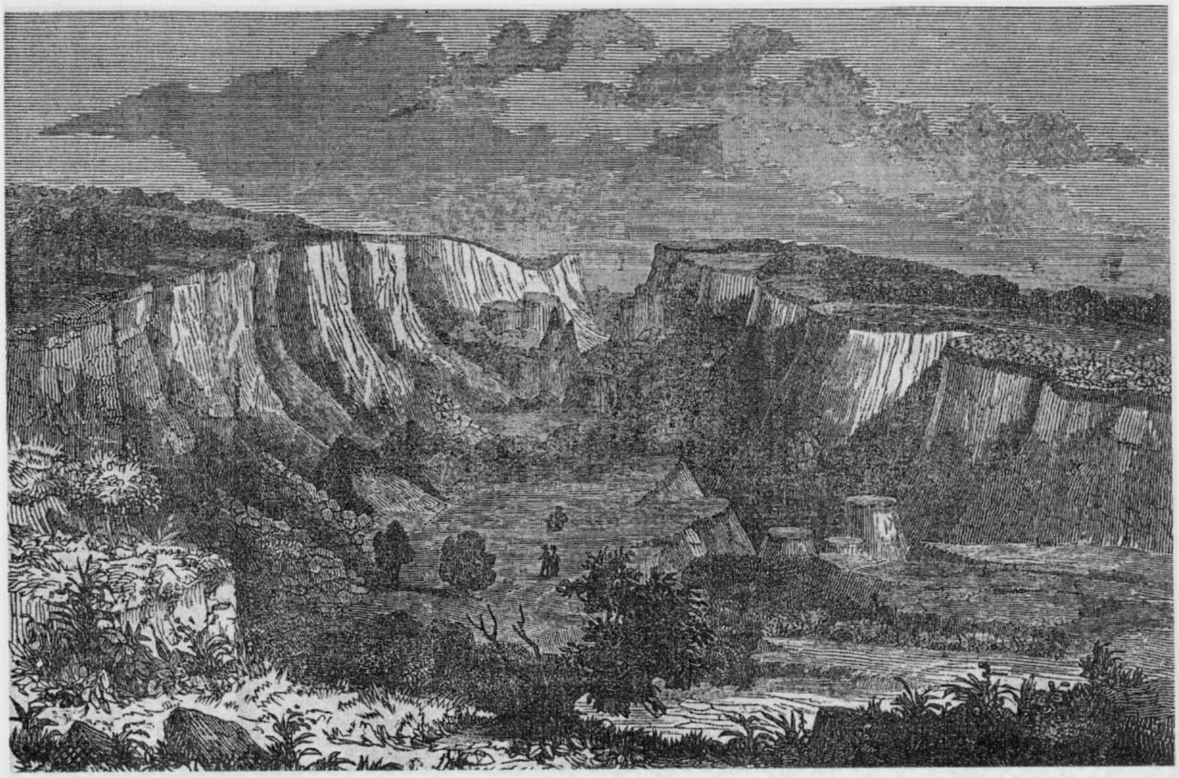


Figure 6.23 A view of 'The Chasm' looking west. From Roberts (1840), reproduced with permission of Lyme Regis Museum.

Figure 6.24 The reef and lagoon at Culverhole Point looking east. An engraving on stone by G. Hawkins Jr, reproduced with permission of Lyme Regis Museum.

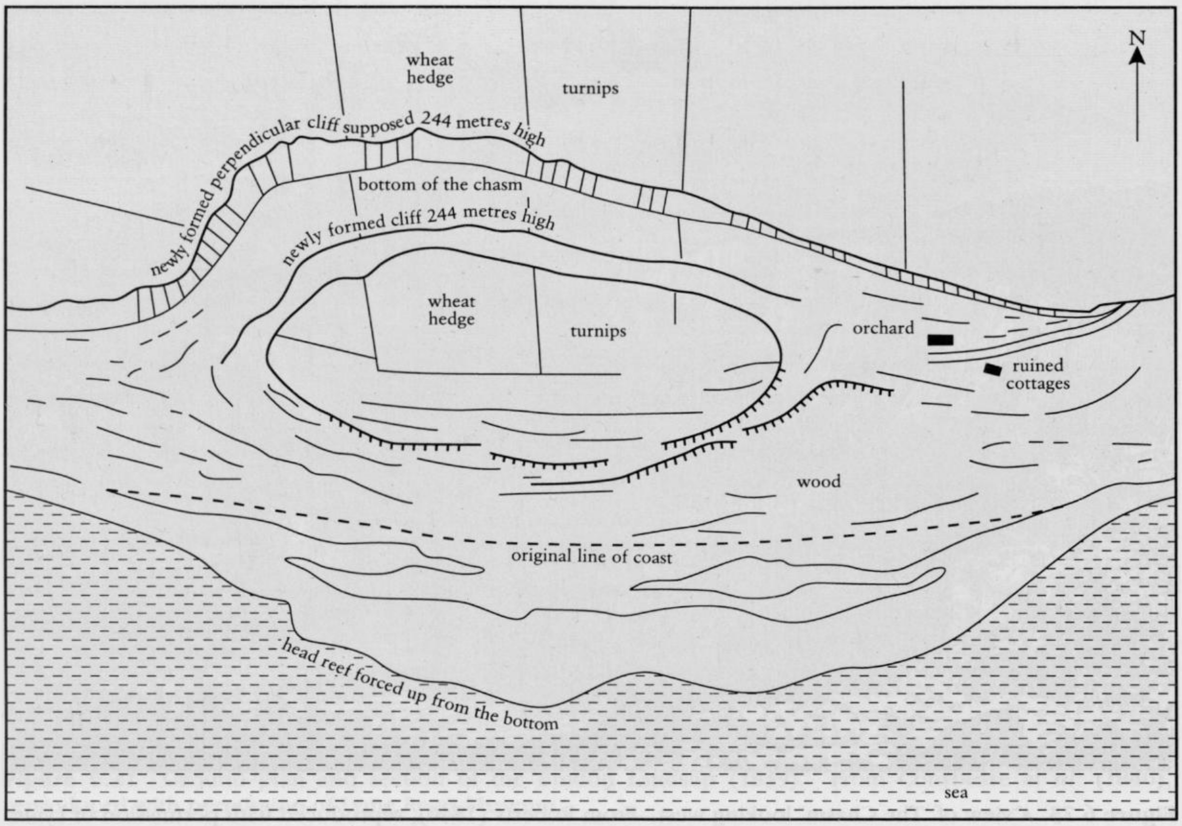


Figure 6.25 Plan of the landslip near Axmouth, Devon. After Anon (1840), from Pitts (1974).

The sea cliffs of the displaced Chalk and Upper Greensand, which had previously stood 15 m to 30 m in height, were now broken and lowered, and thrust 15 m toward the sea. A ridge of the sea shore was pushed up in front of the slip, forming a reef of Upper Greensand. This stretched laterally for nearly 1.2 km, with its outer edge 90 m to 150 m seaward of the previous high-water mark. The beds were much broken, and now dipped inland at angles varying from 30° to 45°, while the surface, which previously had been at least 3 m underwater at low tide, was now raised in places to 12 m above high-water level. The middle of the reef was joined to the mainland by shingle, but one arm extended freely at the western end, and the other to the east enclosed a lagoon which formed a natural harbour. This reef persisted for several years.

Pitts and Brunsden (1987) give a geomorphological map and a geological section of the Bindon slip (Figures 6.26 and 6.27). They made an examination of the groundwater conditions at Bindon, but unfortunately the quality and quantity of groundwater data are very poor for all parts of the site, including Bindon. Only one well exists for which records are available and that is at Dowlands Farm about 1.0 km to the east and 0.5 km inland from the cliff-top. The water level in the well varies very little. Records cited by Roberts (1840) show that the latter half of 1839 was particularly wet, contributing notably to the Bindon slip: at least 50% more rainfall than average was experienced during that period (Arber, 1939).

Pitts and Brunsden (1987) carried out laboratory investigations as follows: samples of the black shales of the Westbury Formation were obtained for determination of residual shear-strength parameters. The samples were obtained from the slipped block at Culverhole Point but were too weathered and disturbed for peak strength determinations to be reliably undertaken.

Residual shear-strength parameters were determined using a 100 mm square shear box. Reconstituted samples of the Westbury Formation shales were formed by consolidation under high loads. A plane was then cut in the

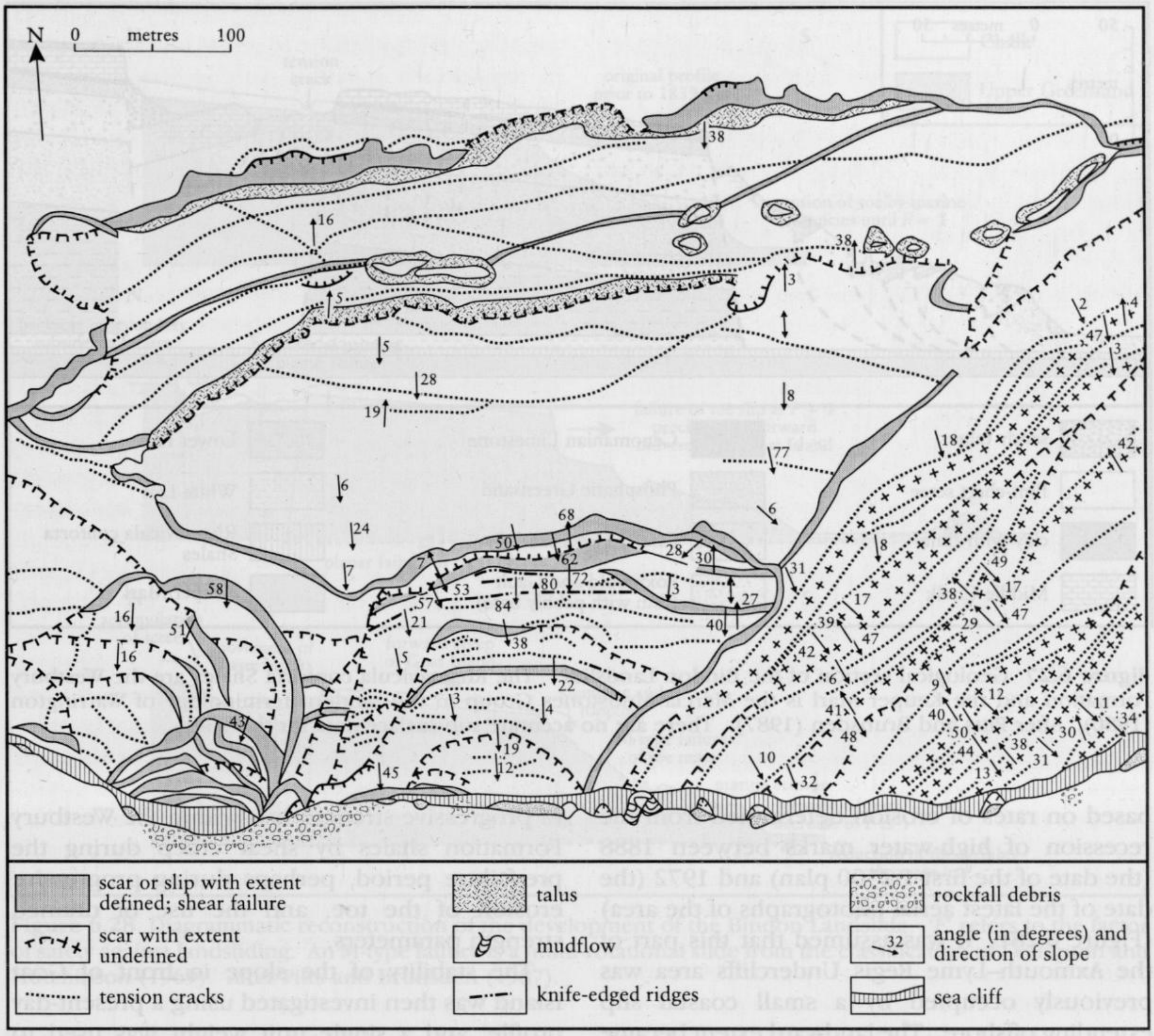


Figure 6.26 Geomorphological map of the Bindon Landslide. After Pitts and Brunsden (1987).

sample and the faces of the cut plane smoothed and polished on a glass plate. The sample was then re-consolidated and sheared several times until a fairly consistent value was obtained. The final pass was made at a low rate of shear. Values of normal stress were increased and the shearing process was repeated. The results obtained for the effective residual cohesion were $c_r = 4\ \mathrm{kN\,m^{-2}}$, and for the effective residual angle of shearing resistance, $\phi = 4.5°$. There must be some doubts as to the validity of such an unusually low value for the angle of shearing resistance, and the rate of shear may have been too great to generate truly drained conditions. Nevertheless, similarly low values were obtained for the Westbury Formation in Somerset by Hawkins and Privett (1985).

Pitts and Brunsden (1987) carried out stability analyses for the 1839 slide at Bindon. The analysis was carried out in terms of effective stresses using the method of Janbu (1973), and Hoek and Bray (1981) for a block slide on a clay layer. Values of pore pressures on the slip-surface were based on water levels recorded in the well at Dowlands Farm. Using considerations detailed by Pitts and Brunsden (1987), this enabled an average value of unit weight, of $20\ \mathrm{kN\,m^{-3}}$, to be used in analysing first-time slides. In analysis of slips incorporating previously slipped material, a 10% reduction in density was assumed.

The slope profile before the 1839 slip was reconstructed using data from contemporary sources, and changes in slope facet positions

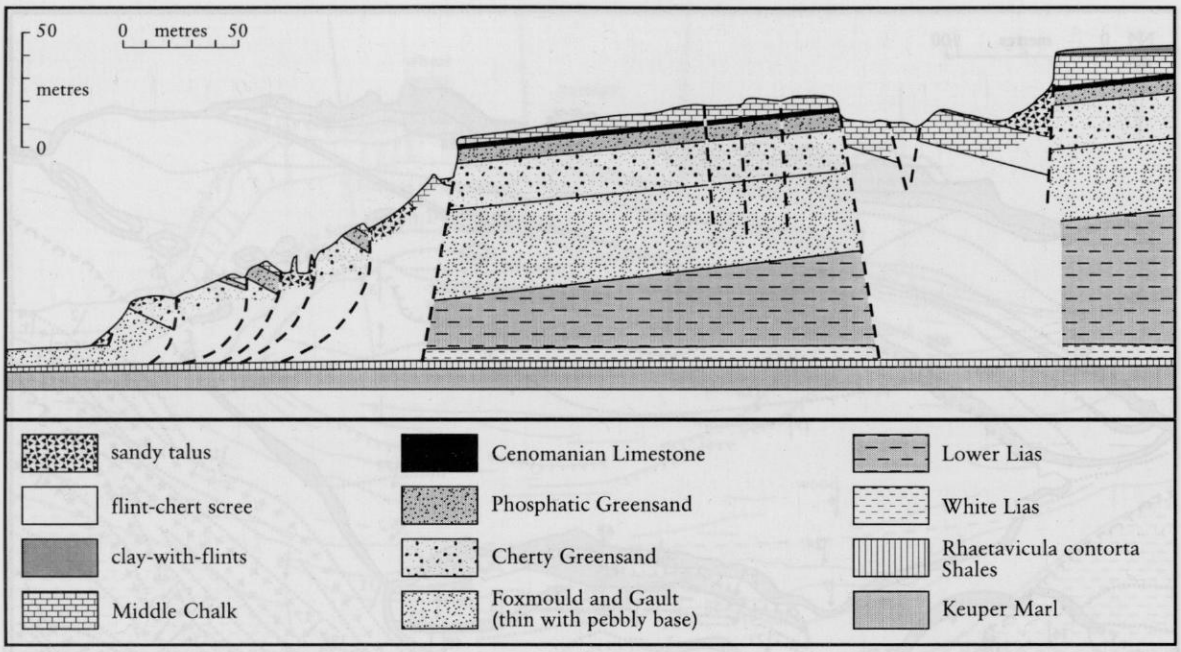


Figure 6.27 Geological section of the Bindon Landslide. The Rhaetavicula contorta Shales are the Westbury Formation and the Keuper Marl is the Mercia Mudstones Group in the modern terminology of Warrington (1980). After Pitts and Brunsden (1987). There are no accurate sub-surface data for this slide.

based on rates of erosion determined from the recession of high-water marks between 1888 (the date of the first 1:2500 plan) and 1972 (the date of the latest aerial photographs of the area) (Figure 6.28). It was assumed that this part of the Axmouth–Lyme Regis Undercliffs area was previously occupied by a small coastal slip extending offshore. The landward extent became the seaward edge of Goat Island in its pre-slipped position. The shear surface was taken to be within the Westbury Formation, and was assumed to develop on bedding partings, a situation in which cohesion values would be very low.

In order to obtain more realistic values of the shear strength of the Westbury Formation, a back-analysis was performed on the simple planar slide which took place at Charton Bay in June 1969 (Pitts, 1986), for which conditions are quite well established. The original profile of this slide was reconstructed from aerial photographs, providing a peak value for the angle of shearing resistance of the bedding of $\phi = 13°$. This value was then used in the analysis for Bindon where peak strengths were required, the cohesion being considered to be zero.

The use of zero cohesive strength throughout the analysis of the failure of Goat Island reflects the probability of sliding along a discontinuity, of progressive strength loss within the Westbury Formation shales by shear creep during the pre-failure period, perhaps during progressive erosion of the toe, and the use of drained strength parameters.

The stability of the slope in front of Goat Island was then investigated using a present-day profile, and a single unit weight was used to represent the slipped material, in conjunction with the residual strength of the Westbury Formation shales along the shear surface. Finally, analyses were undertaken to investigate the state of the stability of the current slope. This considered the three main components of the modern slope: The Chasm, Goat Island, and the slipped mass in front of Goat Island.

A slab slide was assumed despite the existence of some inconsistently back-tilted blocks in the floor of The Chasm. A perfectly planar shear-surface at the base of a deep tension crack sited at the rear of The Chasm was assumed. The results of the back-analysis for the failure of Goat Island and The Chasm produced a value for ϕ' of 20.8° mobilized at failure, a much higher value than that obtained at Charton Bay for the same material. This disparity may relate to the proposed trigger of the failure.

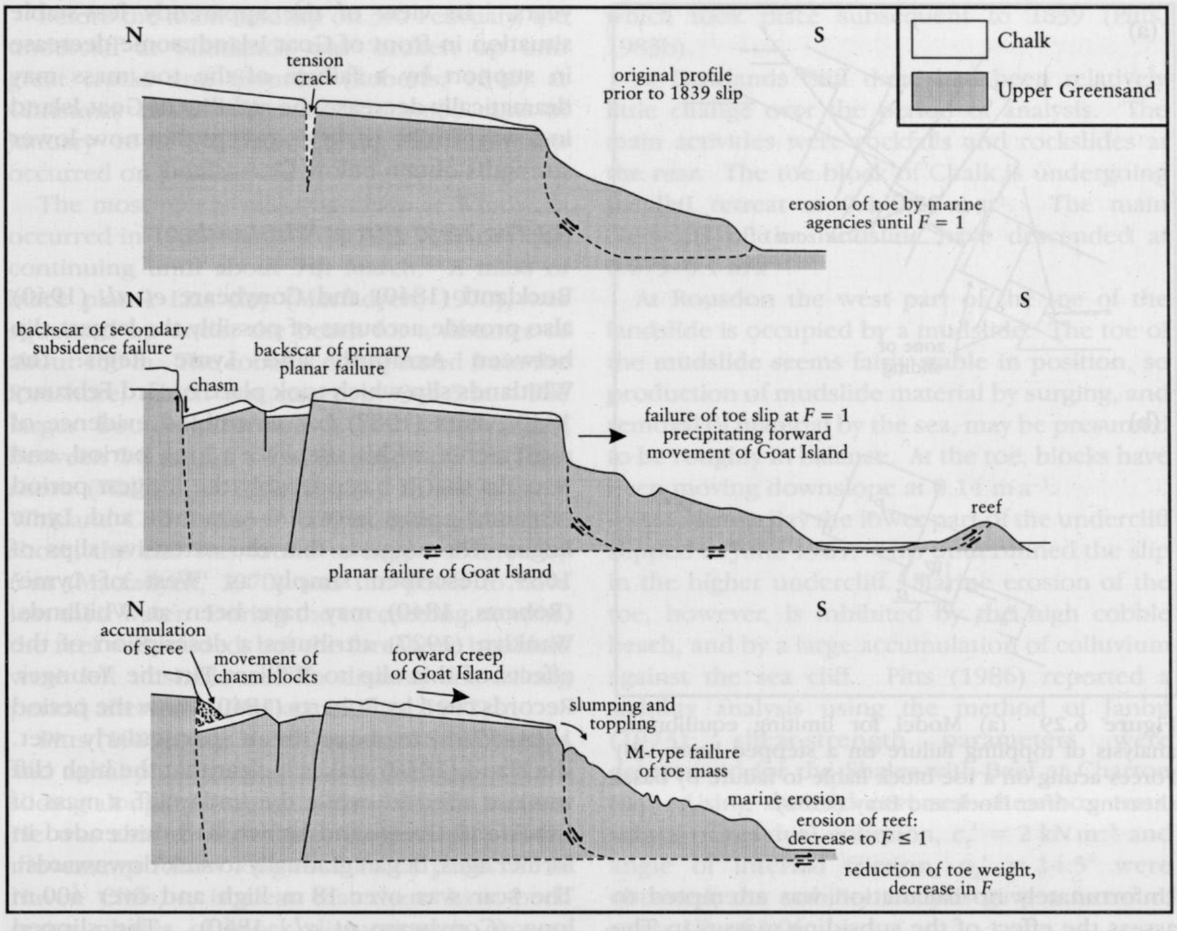


Figure 6.28 Diagrammatic reconstruction of the development of the Bindon Landslide. '*F*' refers to the factor of safety against landsliding. An M-type failure is a multi-rotational slide from the classification of Skempton and Hutchinson (1969). After Pitts and Brunsden (1987).

Several simple stability analyses were carried out to investigate further the order of failure at Bindon. This particularly concerned the hypothesis that the trigger of the main failure of the Goat Island mass was a non-circular rotational failure in front (seaward) of that mass. An analysis of the forces acting on Goat Island when it was unsupported seawards, using the shear-strength parameters from the back-analysis of the Charton Bay slip and the water levels in Dowlands Farm well, in each case produced factors of safety (*F*) of less than 1.0.

An attempt was therefore made to analyse the contribution of the seaward support to the stability of Goat Island. No direct method of analysis seemed to exist that dealt with this contingency. A method was adopted which had been outlined by Hoek and Bray (1981) as a part of a stability analysis procedure for rock masses subject to toppling failure. The formula presented by Hoek and Bray (1981) for calculating the propensity of any of the blocks to slide rather than topple is;

$$P_{n-1} = P_n[W_n(\tan\phi \cos\alpha \sin\alpha)]/(1 - \tan^2\phi)$$

where ϕ is the angle of shearing resistance and the various forces acting on the block are as shown in Figure 6.29. For the situation prior to the 1839 failure, a factor of safety of 1.15 was obtained for Goat Island.

It is difficult to be sure at what stage of failure the toe block was required to be in order to produce a factor of safety (*F*) of 1.0, that is, the factor of safety at which sliding just begins to occur, for Goat Island. The indication is, however, that the failure of the toe mass would have been almost completed before the slip of Goat Island occurred, a factor of safety of 0.99 being obtained for the pre-failure geometry.

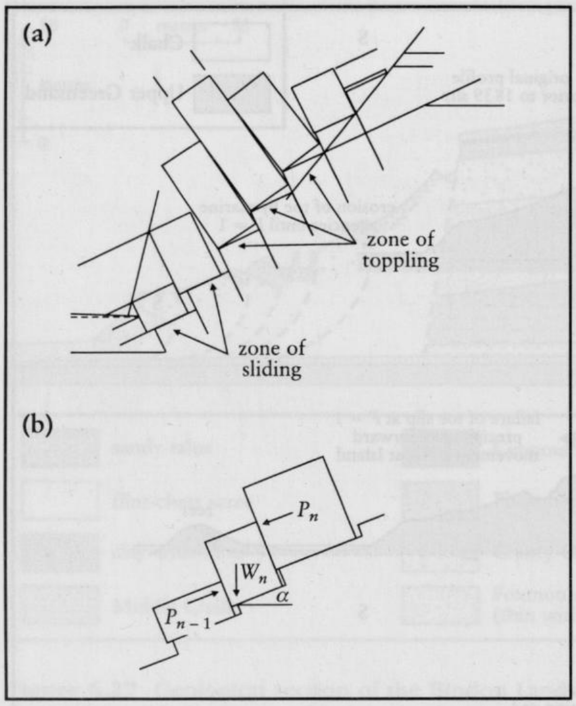


Figure 6.29 (a) Model for limiting equilibrium analysis of toppling failure on a stepped base. (b) Forces acting on a toe block liable to failure by basal shearing. After Hoek and Bray (1981).

Unfortunately no calculation was attempted to assess the effect of the subsiding masses in The Chasm on the stability of Goat Island. Too little is known about the precise course of events to quantify their effects adequately.

An analysis was carried out on the slope geometry to determine the gain in the factor of safety resulting from the lengthening of the profile and the formation of the offshore reef. Although residual strengths operated throughout the length of the failure surface, the factor of safety (F) increased to 1.49.

The slope geometry in front of Goat Island as it exists today has suffered erosion of the reef, about 140 years of marine erosion, and crown loading from degradation of the seaward-facing slope of Goat Island. The factor of safety is very low, around 1.0, and the slope is in a quasi-stable state, if the assumptions made, particularly about shear strengths and groundwater, are realistic.

Finally an analysis was undertaken of the whole slope using the present-day geometry. A value of $F = 1.23$ was obtained, compared to that of $F = 1.49$ for the immediately post-failure situation, a reduction of 17.5% in about 140 years. In view of the apparently less-stable situation in front of Goat Island, some decrease in support by a failure of the toe mass may dramatically decrease the stability of Goat Island in a way similar to 1839, except that now, lower strengths obtain below Goat Island.

(c) The 1840 slip at Whitlands

Buckland (1840) and Conybeare *et al.* (1940) also provide accounts of possibly the largest slip between Axmouth and Lyme Regis: the Whitlands slip which took place on 3rd February 1840. Pitts (1982) has assembled evidence of movements at this site over a long period, and remarks that it has probably the longest period of landslipping between Axmouth and Lyme Regis. He suggests that the extensive slips of 1689, described simply as 'West of Lyme' (Roberts, 1840) may have been at Whitlands. Wanklyn (1927) attributes a description of the effects of this slip to William Pitt the Younger. Records cited by Roberts (1840) show the period 1764–1765 to have been particularly wet. Buckland (1840) makes it clear that the high cliff was not affected, while the undercliff, a mass of Chalk and Greensand 'which had descended in former ages, began gradually to sink downwards'. The scar was over 18 m high and over 400 m long (Conybeare *et al.*, 1840). The slipped material was split into a series of 'irregular ridges and furrows' (Buckland, 1840). Houses on the slipped mass had their floors squeezed upward and their walls were tilted. A nearby garden was converted into a pond of water.

Two reefs close to the shore were seen to 'rise slowly and simultaneously with the slow descent of the subsiding portion of the adjacent undercliff' (Buckland, 1840). This extended about 0.8 km and 30 m seaward of and parallel to the old sea cliff. Buckland (1840) also suggested that 'the bottom of the sea, for a great distance from the present shore, is composed of large fragmentary masses of subsided Chalk and cherty sandstone brought thither by the destructive action of the sea and of land springs in former ages upon ancient undercliffs'. Pitts (1982) points out that this is supported by Conybeare *et al.*'s description of the seaward of the two reefs which was capped by 'a stratum of chalk capped with angular flint gravel, exactly as would be found on the summit of the chalk downs above the undercliff' (Conybeare *et al.*, 1840).

Before the main landslip on 3rd February, the undercliff at Whitlands was 'broken up into great cracks and fissures' (Roberts, 1840) at Christmas 1839. After the 'continued rains of January 1840' (Roberts, 1840) the main slip occurred on 3rd February.

The most recent major landslip at Whitlands occurred in 1981, starting on 28th February and continuing until about 7th March. A mass of black plastic Lias clay (Macfadyen, 1970), was squeezed up under the beach for a distance of about 450 m. The cobbles disappeared from the foreshore, which was raised by 3–4.5 m. The larger boulders formed a soft cliff varying between 1.8 m and 4.5 m in height on the foreshore (Wallace in Pitts, 1982). In the area of Whitland Cliff Pools, 46–229 m inland from the shore, the cliff was found to have subsided by 6 m (Macfadyen, 1970) and the pools to have lost their water. During the succeeding months, the reef was eroded, but the foreshore boulders were left unstable and in disarray for a long period afterwards (Wallace in Pitts, 1982).

Pitts (1983b) has used Ordnance Survey maps and aerial photographs to examine recent landsliding along the Axmouth–Lyme Regis coastline. He was able to show that, in general, present movements follow long-established patterns. At Haven Cliffs, it is clear that there have been major phases of block disruption, a process whereby large blocks are successively broken up into smaller blocks as material moves downslope, as described by Brunsden and Jones (1976). Colluvium has accumulated at the foot of the sea cliff. In the eastern part of Haven Cliffs, this has reduced movement almost to a standstill: scars are degrading and becoming vegetated.

At Culverhole Cliffs, activity has increased since 1905. There are many fresh scars and increased activity around the backscar.

At Bindon, rapid erosion of the reef and the extended toe of 1839 have resulted in steady forward creep of slope elements in front of Goat Island, at an average of 0.3 m a^{-1}, as these elements have been loaded by the degradation of the seaward face of Goat Island by slumping and toppling. As a result, the high-water mark has migrated seawards. Goat Island itself has steadily moved downslope, but the spatial pattern of movement is not uniform: there is a relative lack of movement in the eastern part. This may be due to stabilizing effects of slope movements in the adjacent Dowlands Cliff which took place subsequent to 1839 (Pitts, 1983b).

At Dowlands Cliff there had been relatively little change over the period of analysis. The main activities were rockfalls and rockslides at the rear. The toe block of Chalk is undergoing parallel retreat at 0.1–0.25 m a^{-1}. The main elements of the landslide have descended at 0.075–0.1 m a^{-1}.

At Rousdon the west part of the toe of the landslide is occupied by a mudslide. The toe of the mudslide seems fairly stable in position, so production of mudslide material by surging, and removal of material by the sea, may be presumed to be roughly in balance. At the toe, blocks have been moving downslope at 0.14 m a^{-1}.

At Charton Bay the lower part of the undercliff slipped in June 1969. This undermined the slip in the higher undercliff. Marine erosion of the toe, however, is inhibited by the high cobble beach, and by a large accumulation of colluvium against the sea cliff. Pitts (1986) reported a stability analysis using the method of Janbu (1973): shear-strength parameters were determined for the Shales-with-Beef at Charton Bay. Using drained reversed shear-box tests, values of residual cohesion, $c_r' = 2$ kN m^{-3} and angle of internal friction, $\phi_r' = 14.5°$ were obtained; the stability analysis gave a factor of safety (*F*) of 1.06.

At Humble Point the slope failed in March 1961. There was an apparent advance of the shoreline between 1888 and 1904, which may be an effect of block disruption in the undercliffs, unrecorded elsewhere (Pitts, 1983b). Pitts (1983b) remarks that this slope seems to be in a pattern of evolution towards large-scale slipping. He notes that at Pinhay Bay there was increased mudslide activity in the 1970s, and recession at the toe resulted in 'a substantial re-failure of the face of one of the main blocks within the undercliff. The highly comminuted and weakened nature of much of the slipped material makes its incorporation into the toe mudslide relatively rapid where major seepage from the face of the slope becomes an influence. Major block disruption events at increasingly high upslope positions appear inevitable. The broad, wet, low-angled toe area is being fed by debris which is causing rapid undermining of the upslope blocks. At the same time the crest of the sea cliff continues to recede' (Pitts, 1983b).

Further light is shed on this by Grainger *et al.* (1985; Grainger and Kalaugher, 1995), who report shallow landslide activity that poses a long-term threat to the Hart's Tongue Spring, a large spring of clean water, which issued from the base of a large slipped block of Chalk and Chert Beds, at about 30 m OD and 160 m from the beach, between Whitlands and Pinhay. This source was tapped in 1935 and remains the sole source of water supply to Lyme Regis. Their investigations (survey pegs, temporary bench marks at the pumping station and on large concrete blocks on the beach, continuous-flight auger holes, geomorphological mapping, electrical resistivity and seismic refraction techniques) enabled them to draw a cross-section and reconstruction (Figure 6.30), actual surges of movement being clearly related to rain-

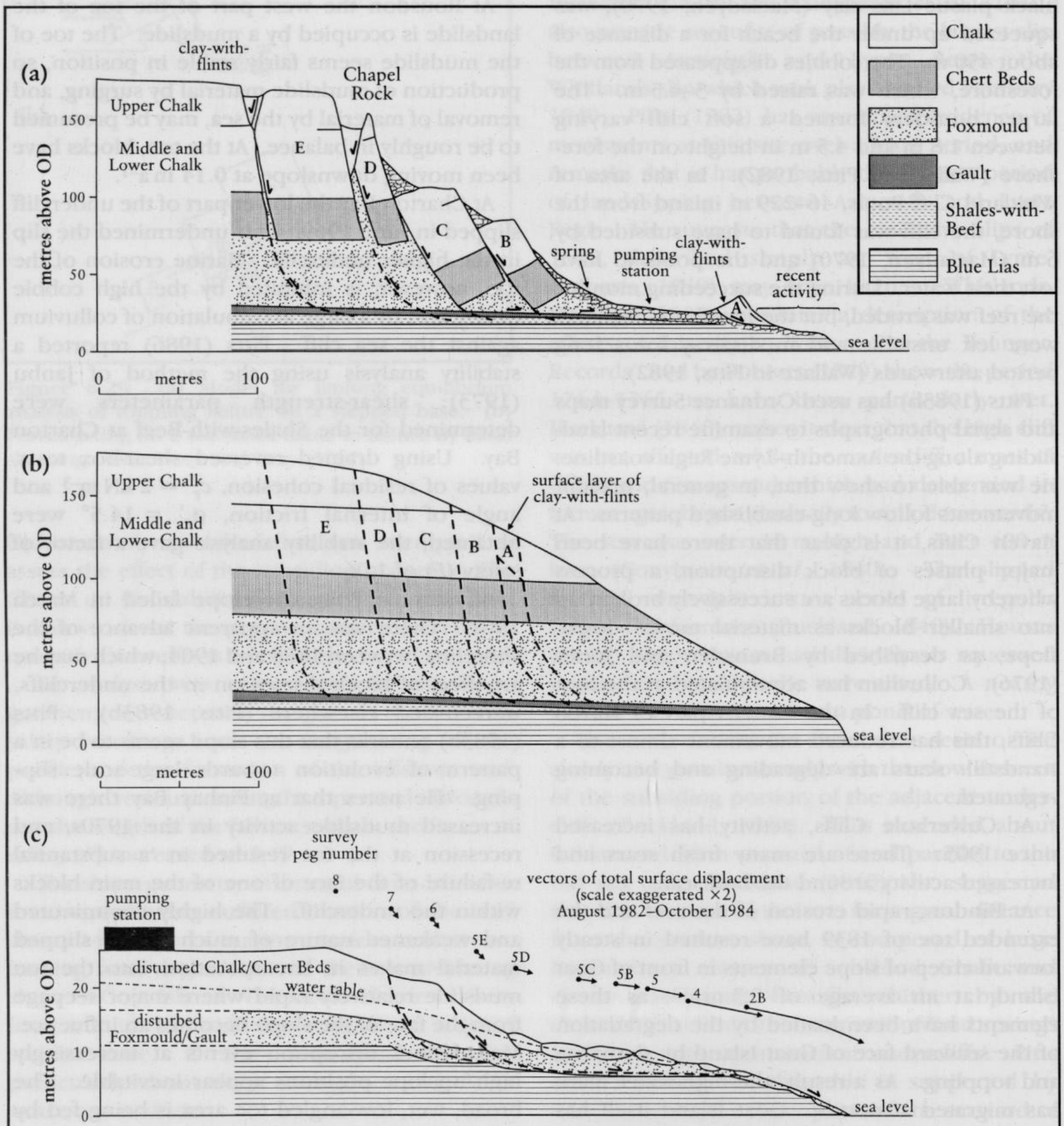


Figure 6.30 Cross-sections and evolutionary reconstruction of the Chapel Rock landslide and the surveyed movements at the undercliff water pumping-station. Note the loss of the Foxmould by flow or extrusion. After Grainger *et al.* (1985).

fall events (Figure 6.31). They were able to conclude that the cliff above the source is currently in a stable condition, but the zone below the pumping station is unstable and unlikely to achieve stability. In this zone retrogressive development of backscarps in the degraded Chalk blocks continues, and at its present rate will reach the pumping station in a few tens of years.

At Ware Cliffs, major block disruption in the lower parts of the slope has been very active since the 1950s. These events have extended the zone of major activity as far back as the public footpath that runs through the Nature Reserve. There is a noticeable boundary between the basically dry and wet parts of the Ware Cliffs slope at a position just downslope of the footpath, from which large amounts of seepage are discharged. Extensive seepage near the toe of the slope and from points in the lower parts of the undercliff are also making the lower slopes unstable. Much material has slipped over the edge of the sea cliff, the effects of which have

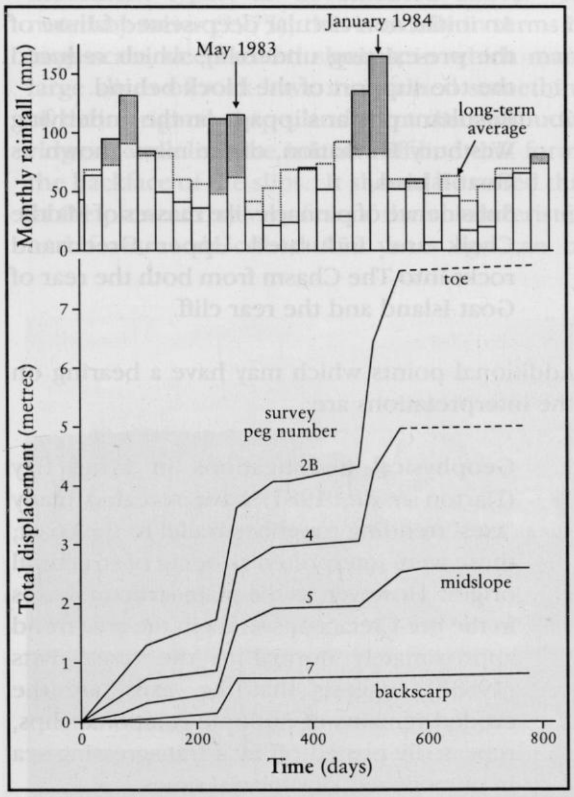


Figure 6.31 Comparison of rainfall and ground movement of the lower slopes of the landslide between Humble Point and Pinhay. After Grainger *et al.* (1985).

been transferred upslope and which have resulted in re-routings of the public footpath to positions progressively farther upslope. This pattern seems likely to continue, resulting in the progressive undermining of the major blocks of Chalk and Greensand in the upper half of the undercliff.

Interpretation

Conybeare (1840; and in Conybeare *et al.*, 1840) and Buckland (1840) suggested that the relatively undisturbed state of Goat Island indicated translational seaward sliding over some saturated horizon. They identified the Foxmould (glauconitic sands, lower division of the Upper Greensand) as the most likely candidate, and postulated that it had been reduced to the condition of a quicksand and 'washed out' by heavy rains, causing undermining of the superincumbent strata, and seaward slipping over the underlying argillaceous Lias and Gault. This is probably the first description of liquefaction and metastable sands in the scientific literature. Arber (1940) suggested that the slip is associated with the Cretaceous overstep, and supported the notion of 'washing away' of the Foxmould leading to undermining and translational sliding. A major (but erroneous) challenge to this view came from Ward (1945) who stated categorically that the main movement was rotational and similar to that at **Folkestone Warren**. This was questioned by Arber (1962), in view of the relatively undisturbed nature of Goat Island, with its (if anything) slightly seaward tilt. The strongest expression of this view is a diagram by Macfadyen (1970; Figure 6.32). However, a close examination of a series of ridges running parallel to the edge of Goat Island on its seaward side unfortunately led her to a later acceptance of Ward's view (Arber, 1971). Arber has since pointed out (1973) that the back-tilting of blocks in The Chasm, and the pushing up of Lower Greensand strata in the reef, support this view, and has suggested that Goat Island may not have moved at all. However, morphological mapping by Pitts (1979) suggests that some foundering and seaward slipping have in fact taken place. It is now widely recognized that the mechanism of 'graben' or 'chasm' formation requires forward movement of the foundered blocks and is a diagnostic feature of non-circular failure. The Macfadyen (1970) model should be discarded.

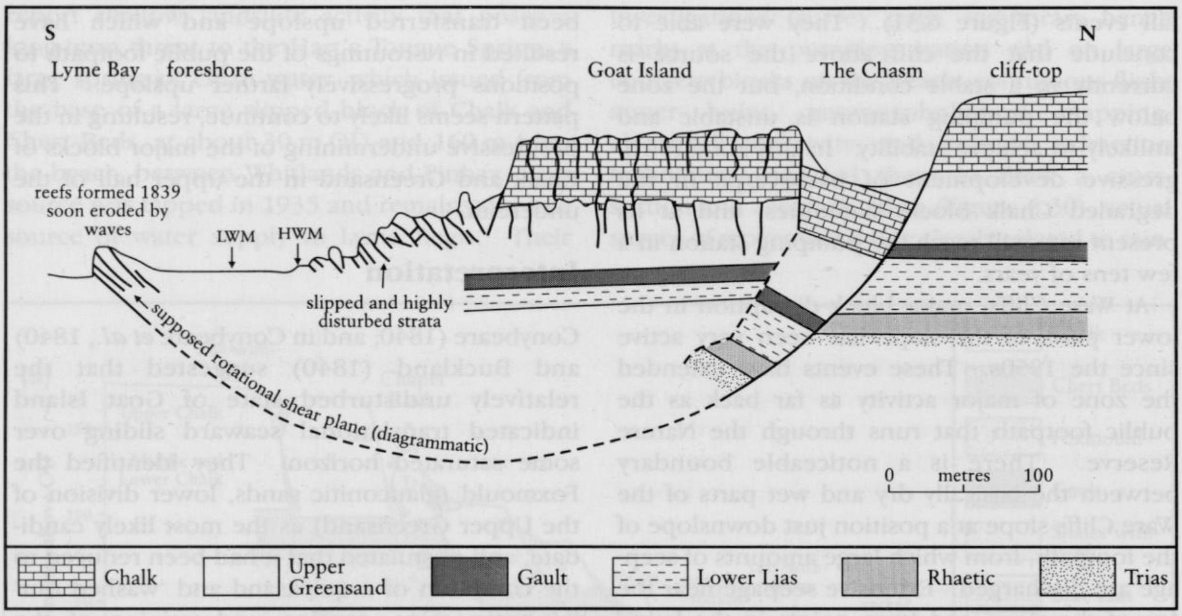


Figure 6.32 Hypothetical section through the Bindon Landslide showing a rotational failure mechanism (after Macfadyen, 1970, from Pitts, 1974). The model is not substantiated by sub-surface information. Note that the model does not explain the toe slips, nor why the strata in Goat Island remain horizontal when subject to rotational movement. The graben is diagnostic of a non-circular failure on the bedding.

Pitts (1983b) concludes from his study of recent movements that present-day developments depend primarily on conditions at the toe. The removal of the toe of a landslide, landslide debris, slipped blocks, or even offshore reefs subsequent to a slope failure or the gradual recession of sea cliffs upon which landslides have developed, result in critical fore-shortening of the profiles and a reduction in the factor of safety. Progressive downslope movement of the large, more-competent slipped blocks behind then takes place more easily. Processes at the rear, notably rockfalls from the backing cliffs (rear scarp), and the general accumulation of scree at their foot seem to result in crown loading but this appears to have only a marginal influence on the destabilization of the remainder of the slope. As Pitts (1983b) remarks, this is borne out by the rarity of whole-slope failures which are not demonstrably preceded by a series of increasingly large failures developing rearwards from the toe. He concludes by observing that over the time period which map, aerial photograph and field observation cover, there seems little reason to believe that instability following the currently established patterns will not continue.

With respect to the 1839 Bindon landslide, Pitts and Brunsden (1987) propose that the mechanisms were as follows:

1. An initial non-circular deep-seated failure of the pre-existing undercliff, which reduced the toe support of the block behind.
2. Resultant planar slippage on the underlying Westbury Formation, of the mass known as Goat Island.
3. Subsidence of pinnacle-like masses of Middle Chalk and indurated Upper Greensand rocks into The Chasm from both the rear of Goat Island and the rear cliff.

Additional points which may have a bearing on the interpretations are:

1. Geophysical investigations in Lyme Bay (Darton *et al.*, 1981) have revealed many 'axes' trending roughly parallel to the coast; these were interpreted as being of structural origin. However, as the main structural axes in the pre-Cretaceous strata in the area trend approximately normal to the coast, Pitts (1983b) suggests that the 'axes' are the eroded remains of multiple rotational slips, repeatedly planed off by a transgressing sea in immediately post-glacial times.
2. Alternatively (Kellaway *et al.*, 1975) the landslips may have resulted from glacial oversteepening and dissection by meltwaters, but evidence for glaciation is inconclusive.

Conclusions

The Axmouth–Lyme Regis site is worthy of GCR status for several reasons, and has additional historical importance. Owing to the extraordinary co-incidence of two eminent geologists (Conybeare and Buckland) being in the vicinity at the time of the major 1839 landslip, the site was meticulously examined and recorded at the time, the first occasion that this had happened with any landslip, worldwide. The results of the slip were scenically so spectacular that many pictures of it were drawn and painted and accounts of the events collected. Conybeare's explanation represents an important landmark in the development of attempts to explain and understand mass movements. Secondly, the existence of Goat Island beyond The Chasm at Bindon has led to an instructive controversy over the nature of the 1839 slip. Thirdly, the entire stretch of cliffs (the Ware Cliffs slides are currently extending eastwards into the gardens of Lyme Regis), shows an astonishing and quite exceptional variety and richness of mass-movement types, as documented above. As noted by Arber (1973), this is chiefly in terms of large-scale, deep-seated slumps, in which many large slices of rock have maintained something of their original shape, and form discontinuous ridges paralleling the inland cliff-line that forms the backface of the slips. It should be noted that at the present day very few of the characteristics and features described 150 years ago can be observed with ease at the site, as the vegetation cover is very thick and in many places virtually or actually impenetrable.

BLACK VEN, DORSET (SY 347 927–SY 363 931)

R.G. Cooper

Introduction

The Jurassic outcrop in Great Britain terminates on the south coast in Dorset, where some of the finest landslides in Britain are located. At many locations along the Dorset coast, a permeable caprock overlies an impermeable clay. The structural relationships of the Jurassic strata are such that a variety of different subjacent beds form caprock-and-aquiclude pairs, and within this range of landslide generators, each has different properties and styles of failure (Brunsden, 1996b).

Among these, Black Ven (Figures 6.33 and 6.34), located between Charmouth and Lyme Regis, is of particular interest because of its active mudslides. Indeed, it is the most active and complex landslide site in the British Isles. It comprises rotational slides, topples, rockfalls and slumps in Upper Greensand, above mud-slides, mudflows and sandflows, which feed down to the beach across Liassic materials. It has a long history (Lang, 1928; Arber, 1941,

Figure 6.33 The Black Ven landslide. (Photo: R. Edmonds, Dorset County Council.)

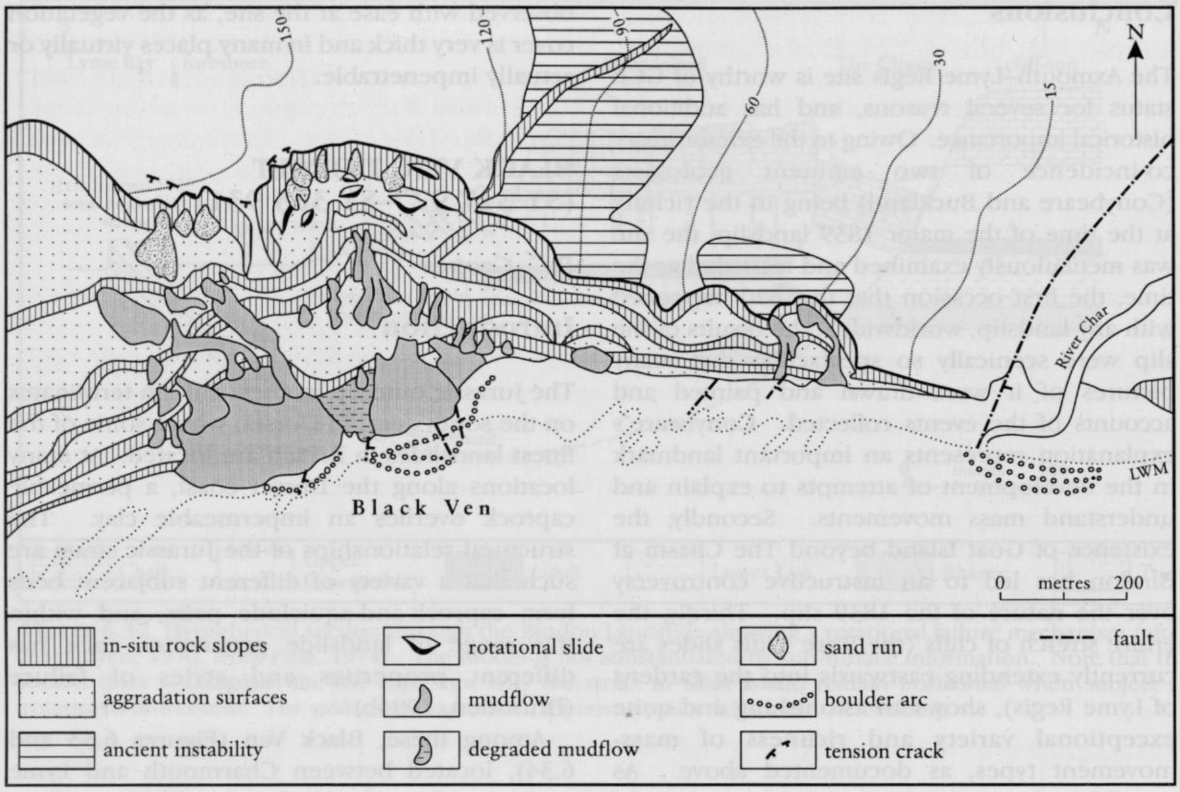


Figure 6.34 The mass-movement complex at Black Ven as it appeared in 1974. After Conway (1974).

1973; Wilson *et al*., 1958; Brunsden, 1969, 1984; Brunsden and Goudie, 1981; Chandler and Cooper, 1988, 1989; Chandler and Brunsden, 1995; Brunsden and Chandler, 1996), but its present character was established by major movements that took place in 1956–1957, 1958 (Conway, 1974) and more recently by renewed activity in the 1980s and 1990s (Figure 6.35). The cliffs reach 150 m above OD. Cliff retreat in the order of 5 m a^{-1} to 30 m a^{-1} is typical during periods of activity. Between major events, erosion at the toe is around 15 m a^{-1} to 40 m a^{-1}. During periods when detached material at the head of the slope is highly saturated, debris tumbles off the edge of an upper bench and drops 20 m onto the middle of three terraces. The upper segment of the Black Ven slope is therefore loaded at the head of each bench. The change begins at the rear scar, where the cycle of primary instability is generated. The original road along the coast was destroyed by landslips in the 18th century (Koh, 1990). A cart track running parallel to the road 100 m farther inland disappeared in 1965 and a section of the Heritage Coast Path collapsed in 1985; it was renewed, but lost again in 1994.

By the time the moving material has reached the middle terrace, mudslides and mudflows have developed. The debris is funnelled into large mudslide tracks, which pour across the cliff separating the middle and lower terraces. The process repeats as the material moves across the top of the Blue Lias towards the beach where the mudslides merge to form large composite fans and toe lobes.

Description

The Black Ven cliff is composed of the Blue Lias, Shales-with-Beef, Black Ven Marls and Belemnite Marl divisions of the Lower Lias (Lower Jurassic), overlain unconformably by Gault Clay and then by the Foxmould and Chert Beds divisions of the Upper Greensand (Lower Cretaceous) (Figure 6.36).

Dips in the Jurassic beds are about 2°–3° south-east or ESE, and the plane of unconformity at the base of the Cretaceous strata dips 1°–2° south or SSW. The cliff profile (Figure 6.37) shows well-developed terraces at the levels of the base of the Black Ven Marls (Conway,

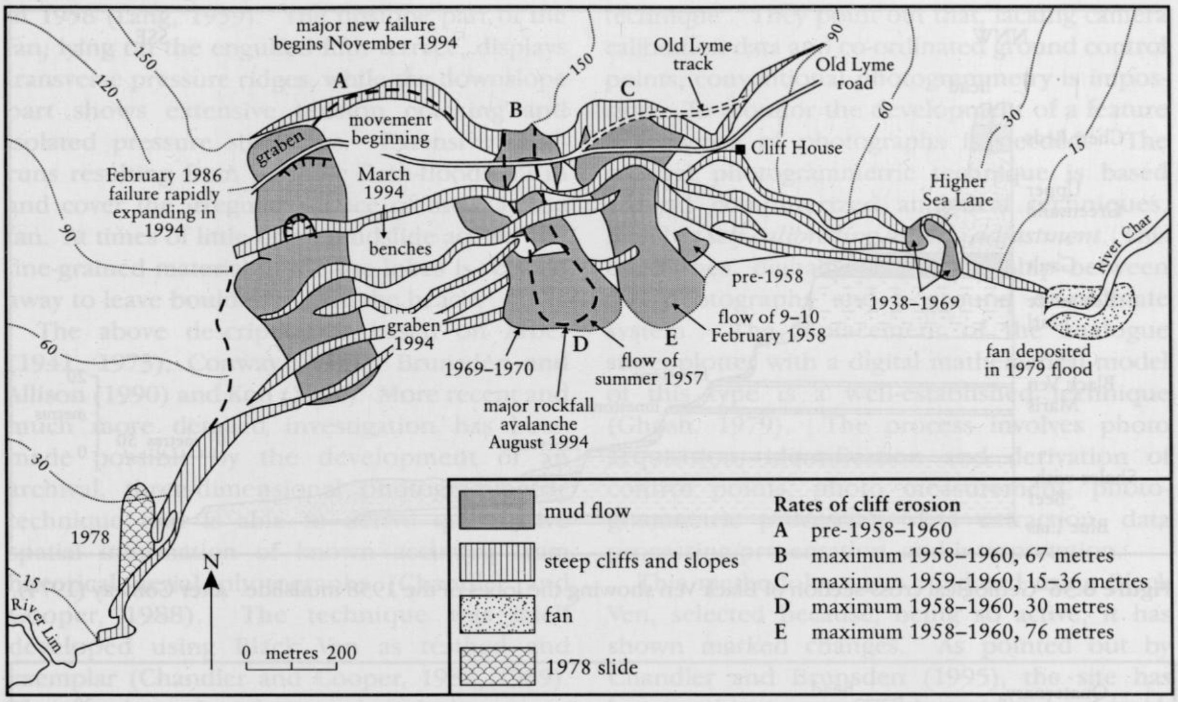


Figure 6.35 The Black Ven mudslide complex showing movements between 1958 and 1994. After Chandler and Brunsden (1995).

1974). These terraces or benches are caused by the presence of resistant horizons within the Lower Lias. A fourth resistant horizon gives rise to a less well-developed terrace feature at the top of the Blue Lias, but this is often obscured by landslip deposits. A number of minor terraces are developed above the other resistant horizons, but they are not very extensive horizontally.

In the Upper Greensand, brecciated Chert Beds consist of broken chert in a firm, coarse sandy clay matrix with some iron and manganese oxide concentration in the lower part. The beds are much harder than the underlying decalcified Foxmould sands and this has resulted in the development of a steep upper cliff, the height of which is sufficient to allow the generation of shear stresses far in excess of the resistance offered by the decalcified sands. This results in the propagation of single- and multiple-failure rotational slides that affect the full thickness of the Upper Greensand. Secondary iron oxides have been deposited at the base of the Chert Beds, impeding the downward movement of groundwater. This results in springs being thrown out at the cliff-face, which have cut deep gullies in the cliffs of Foxmould sand below. This process is greatly assisted by land drainage from the top of the cliffs. Many of the initial movements of slide blocks occur as a result of failure of the conical buttresses that develop between these gullies.

The base level to which the gullying and the rotational slides operate could be the top of the Gault Clay, the level of the highest terrace. Large accumulations of sand and chert debris build up on this terrace, and during the winter months are rapidly saturated. The water is discharged on the cliff-face at the junction of the debris with the Gault Clay. This results in extensive seepage erosion (Conway, 1974) and gullying, leading to failure, and the debris is carried down the gullies onto the next terrace below. The upper cliff sides are thus deprived of part of their toe areas and stress is again able to build up to the level required to regenerate the cycle of primary instability.

Although gullied by seepage erosion the major cause of the removal of material from the Belemnite Marl cliffs are joint-controlled rockfall caused by erosional unloading, and frost action. The material received by the second terrace, at the base of the Belemnite Marl cliffs, is again rapidly saturated in winter and loaded at its head by the cascade of mud from the terrace above. It discharges water and sediment onto

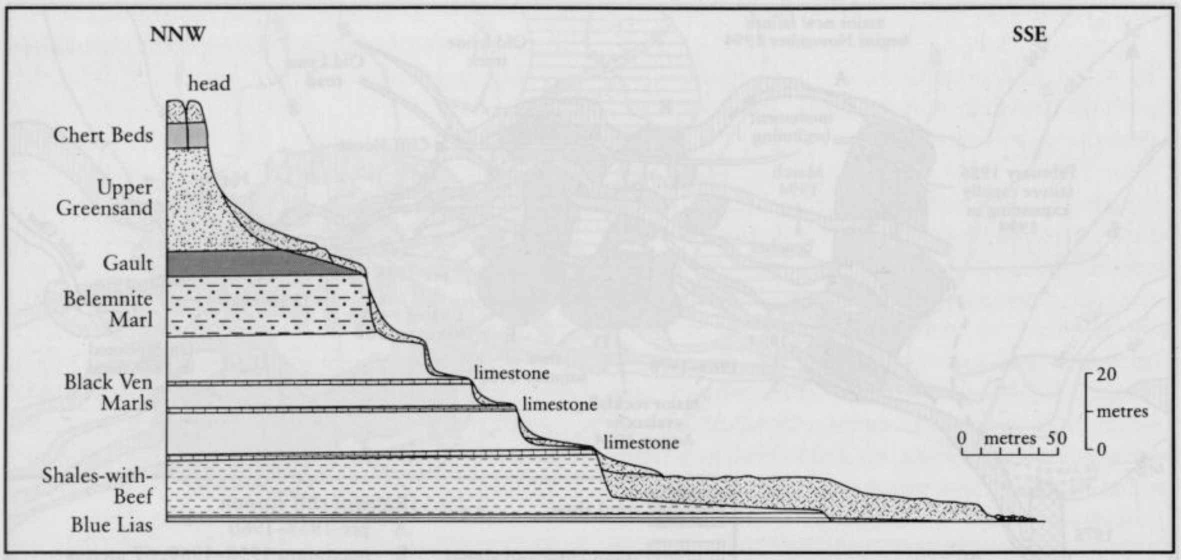


Figure 6.36 Geological cross-section of Black Ven showing the lobes of the 1958 mudslide. After Conway (1974).

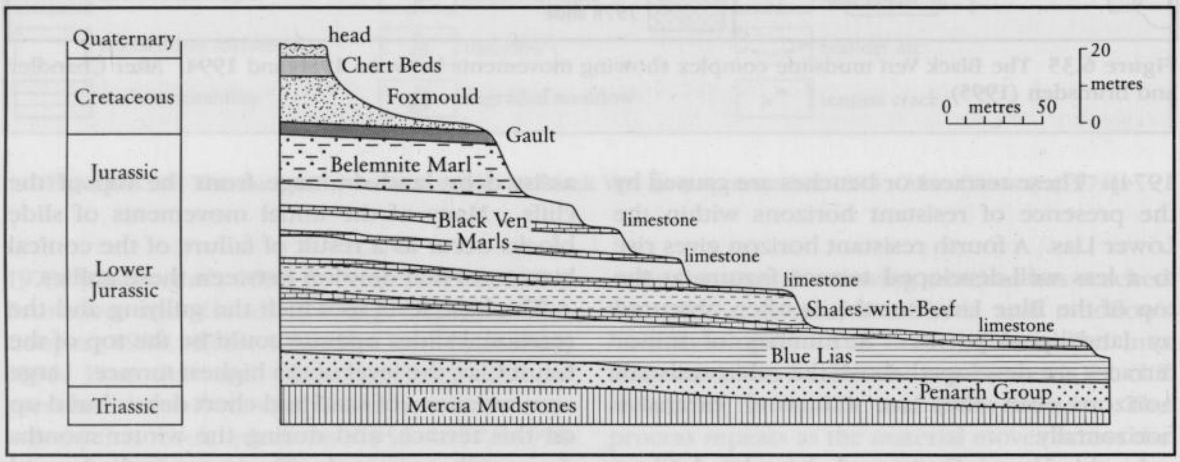


Figure 6.37 Section through cliffs to the west of Black Ven showing regional dip and benches under-scoured by landslide debris.

the next cliff-face from its lower boundary. The Black Ven Marls below is soft and fissured. Subjected to seepage erosion, it is rapidly gullied and small rotational failures occur. Water-charged debris, carried down gullies or pushed over the terrace edge by material accumulating behind, strips off the outer, weathered, layer of clay. Removal of this material from the terrace face results in steepening of the cliff and allows the generation of higher stresses in the clay, which in consequence approaches a failure condition. The clay, which is heavily over-consolidated, takes in water, swelling and weakening until failure takes place. On the benches major mudflows, mudslides and sheet-flows descend to the third and fourth terraces at the base of the Black Ven Marls.

The process is repeated from the fourth terrace over the soft Shales-with-Beef cliff down to the beach. Flows and slides coalesce to form large composite toe fans, which often (as in some recent years) completely envelop the fifth and lowest terrace, at the top of the Blue Lias, and much of the beach. This terrace is really a toe fan, a composite body formed by the accumulation of debris from many cycles of secondary instability, and built on an original which resulted from a single catastrophic event in 1958 (Lang, 1959). The upslope part of the fan, lying on the engulfed fifth terrace, displays transverse pressure ridges, while the downslope part shows extensive tension cracking and isolated pressure structures. Extensive sand-runs resulting from periodic flash-floods fill in and cover the irregular surface of the clay toe fan. At times of little or no mudslide activity the fine-grained material in the toe lobes is washed away to leave boulder arcs on the beach.

The above description is based on Arber (1941, 1973), Conway (1974), Brunsden and Allison (1990) and Koh (1990). More recent and much more detailed investigation has been made possible by the development of an archival, three-dimensional photogrammetric technique that is able to derive quantitative spatial information of known accuracy, from historical aerial photographs (Chandler and Cooper, 1988). The technique was itself developed using Black Ven as testbed and exemplar (Chandler and Cooper, 1988, 1989). These authors show how analytical photogrammetry can be applied to historical photographs, a hitherto untapped source of data for geomorphologists and other Earth scientists. They term their research the 'archival photo-grammetric technique'. They point out that, lacking camera calibration data and co-ordinated ground control points, conventional photogrammetry is impossible. To monitor the development of a feature a sequence of photographs is needed. The archival photogrammetric technique is based around computerized analytical techniques, mainly a *self calibrating bundle adjustment*. This establishes, digitally, the relationship between the photographs and a ground co-ordinate system. The replacement of the analogue stereoplotter with a digital mathematical model of this type is a well-established technique (Ghosh, 1979). The process involves photo acquisition, identification and derivation of control points, photo measurement, photogrammetric processing, data extraction, data processing/presentation, and interpretation.

This methodology was validated using Black Ven, selected because, being so active, it has shown marked changes. As pointed out by Chandler and Brunsden (1995), the site has been subject to several aerial photograph surveys constituting the 1946, 1958, 1969, 1976 and 1988 aerial photographic 'epochs' (see Figures 6.38–6.42). A further epoch, for 1995, is analysed by Brunsden and Chandler (1996).

Figure 6.38 Aerial photograph for the 1946 epoch. (Photo: English Heritage (NMR) RAF Photography.)

Figure 6.39 Oblique aerial photograph for the 1958 epoch. (Photo: Crown Copyright/MOD. Reproduced with the permission of the Controller of Her Majesty's Stationery Office.)

Figure 6.40 Aerial photograph for the 1969 epoch. (Photo: Copyright reserved Cambridge University Collection of Air Photographs.)

Figure 6.41 Aerial photograph for the 1976 epoch. (Photo: reproduced by permission of Ordnance Survey on behalf of HMSO © Crown Copyright (2006). All rights reserved. Ordnance Survey Licence number 100038718.)

Figure 6.42 Oblique aerial photograph for the 1988 epoch. (Photo: J. Chandler.)

The basic data units used for all photogrammetrically based methods are the three-dimensional co-ordinates that can be obtained with a density and efficiency which is unobtainable by other techniques. Data extraction is greater, denser and with subsequent data processing, more powerful and flexible. The co-ordinate data can be used to provide basic planimetry, slope profiles and contours (Chandler *et al.*, 1987) (see Figure 6.43), digital terrain models (DTMs) (shown as isometric views in Figure 6.44) and movement vectors. The technique can be regarded as an updating of previous methods of geomorphological mapping. As remarked by Chandler and Cooper (1988), precise definition and coding of morphological boundaries by rigorous photogrammetric techniques combines the benefits of geomorphological interpretation with positional relevance. Visual comparison between photographs at two widely differing times provides a basic tool which can be used to identify, quantify and interpret areas that display any degree of change.

Although contour plans provide a full description of site morphology at the different epochs it is difficult to identify areas of change by visual inspection. However, subtracting a grid surface produced at one epoch from the grid of a later or earlier epoch creates a grid surface that represents the change of form over the period defined by the photographs. This surface can be contoured, thus quantifying the spatial effects of processes: some areas will have lost material, others will have gained material, and some will have exhibited no change. Chandler and Cooper (1988) caution that the last-mentioned set of areas are not necessarily inactive areas. They can be areas where the input of material has equalled output over the defined period (see Figure 6.45a–f).

Chandler and Brunsden (1995) deal in more detail with the problems of applying photogrammetric methods to archive photographs, in particular the components of the self-calibrating bundle adjustment. At Black Ven the control points used at all epochs are derived from one Ordnance Survey plan, which therefore acts as a datum.

Koh (1990) set up an automated data-gathering system recording rainfall, porewater

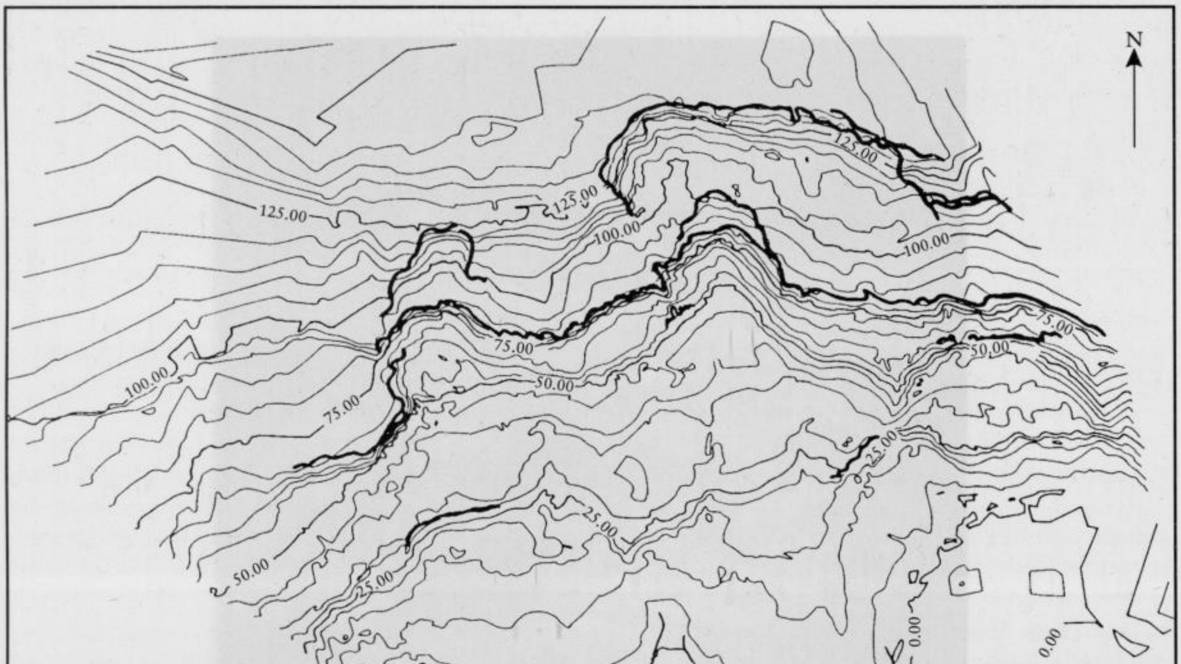


Figure 6.43 Contour plot of Black Ven after the 1958 movements produced by interpolation of 11 000 data points established by photogrammetry. After Chandler and Brunsden (1995).

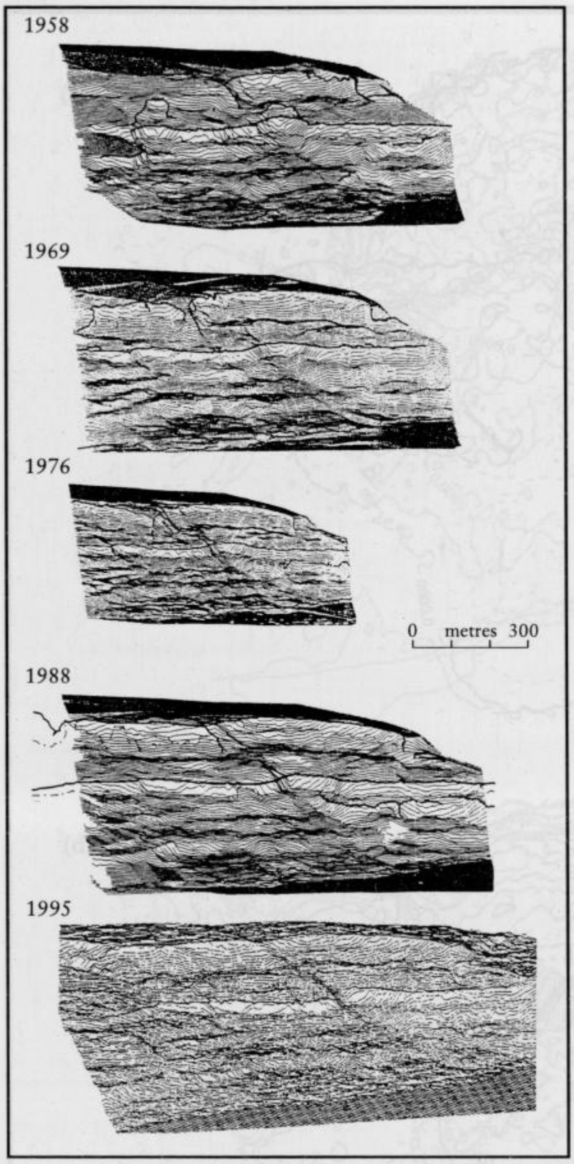


Figure 6.44 Digital terrain models (DTMs) shown as isometric views for the Black Ven mudslide at five epochs between 1958 and 1995. Note that the 1958–1988 epochs are based on analytical photogrammetry and an 11 000 point data set. The 1995 model is based on a larger data set aquired by digital photogrammetry. After Chandler and Brunsden (1995).

pressure, loading, surface movement and subsurface displacement, below the Belemnite Marl cliffs at Black Ven. His results are summarized in Figure 6.46 where the response of porewater pressure and cumulative displacement to monthly rainfall is very apparent. Koh suggests two alternative mechanisms: visco-plastic flow as described by the Bingham equation, and the release of dissolved solids, according to the groundwater chemistry shear-strength model of Moore (1988).

Interpretation

(a) 'The reservoir principle'

It has been suggested by Denness (1972) and Conway (1974) that the sequential process active at Black Ven may best be understood in terms of the presence, in intimate association with the instability, of bodies of material that behave as reservoirs of groundwater with effectively impermeable floors. Naturally occurring groundwater reservoirs may be seen as consisting of two kinds, primary and secondary, depending on whether the reservoir material itself is an in-situ rock body or an accumulation of rock debris resulting from slope degradation. Hence, the in-situ Upper Greensand is the primary reservoir at Black Ven. The debris accumulations on the terraces below are secondary reservoirs, and it is the gradual release of accumulated water from these that leads to the unusually rapid degradation of the material on the terraces and the rapid transport of their material to the cliff-foot.

(b) The episodic landform change model of Chandler and Brunsden (1995)

Chandler and Brunsden (1995) include a 'Speculative Discussion' based on results from the archival photogrammetric technique. One view of morphological change is that landform change takes place when a state of process equilibrium and morphological stability is perturbed by an impulse of change of sufficient character to overcome the tolerance of the system (Brunsden and Thornes, 1979; Brunsden, 1985, 1990). This overcoming of tolerance may be divided into two phases: 'preparatory' impulses, which predispose a system to change, and 'triggering' impulses, which actually push the system over a threshold. In the case of Black Ven there is evidence that the system was prepared for a new phase of mudsliding by the erosion and steepening of the cliffs to a new average angle exceeding 19°. The 1958–1959 mudslides failed at about 19°, which can be regarded as a failure threshold. However, this is in part an

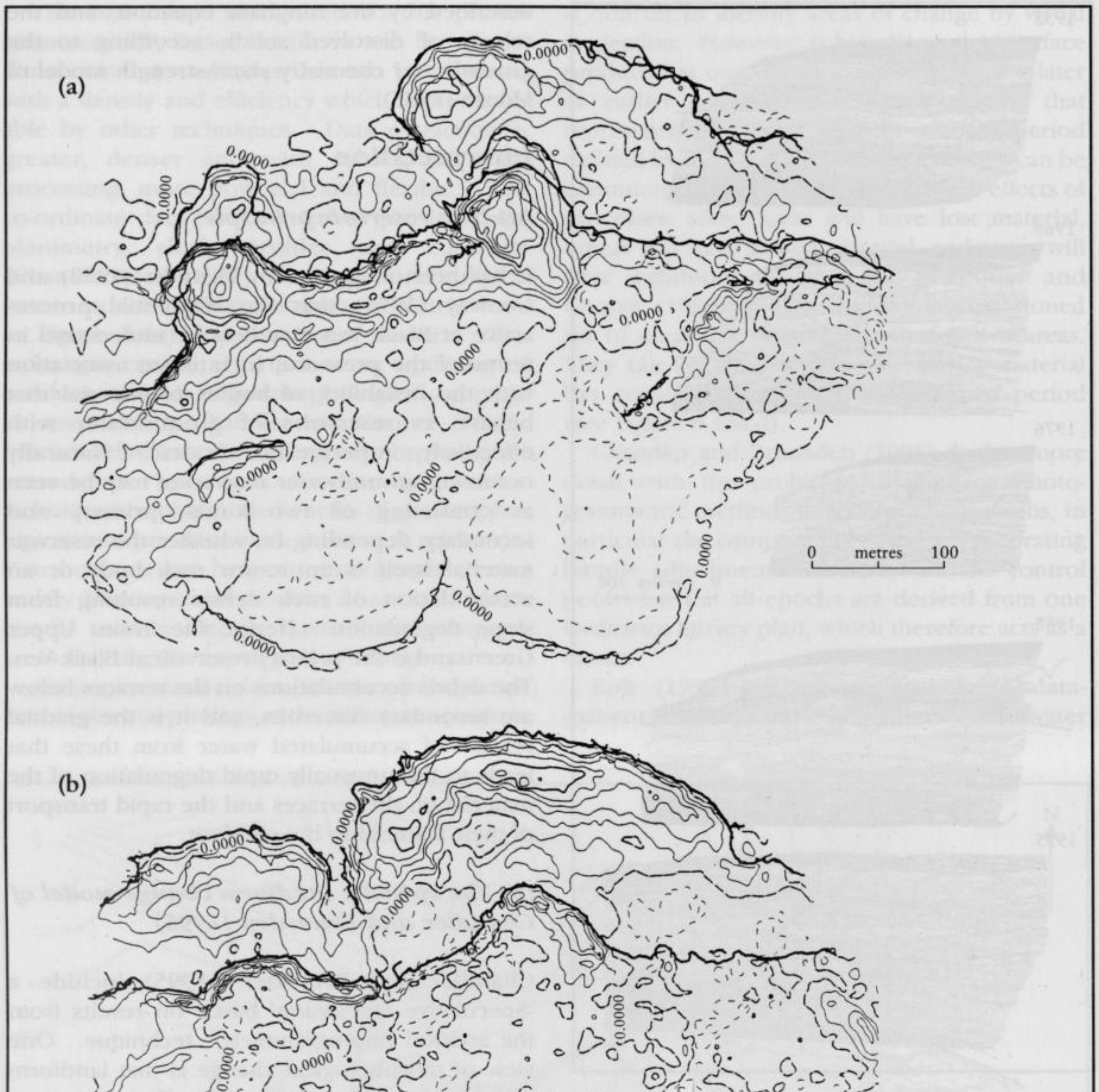


Figure 6.45 Contours of surface difference in elevation (i.e. erosion–deposition–no change) for the periods (a) 1958–1946, and (b) 1969–1958. Period (a) shows the location of the 1958 failures. This can be regarded as the formative event. Period (b) shows the diffusion of the wave erosion of the toe and continued input. The 'no change' along the main mudslide axis shows input = output and dynamic equilibrium over a decade interval. After Chandler and Brunsden (1995). *Continued opposite.*

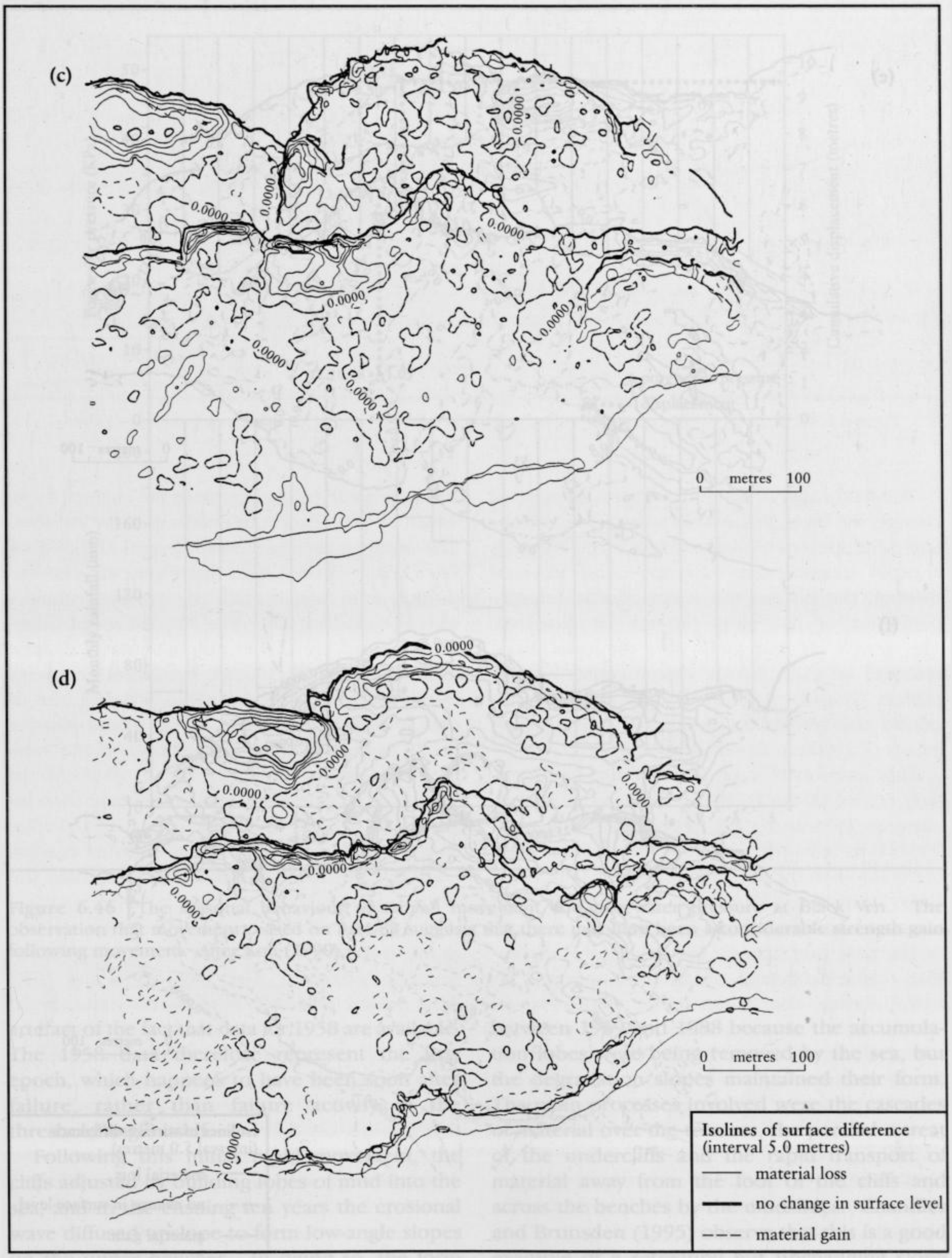


Figure 6.45 – *continued*. Contours of surface difference in elevation (i.e. erosion–deposition–no change) for the periods (c) 1976–1969, and (d) 1988–1976. The 'no change' along the main mudslide axis shows input = output and dynamic equilibrium over a decade interval. After Chandler and Brunsden (1995). *Contined overleaf.*

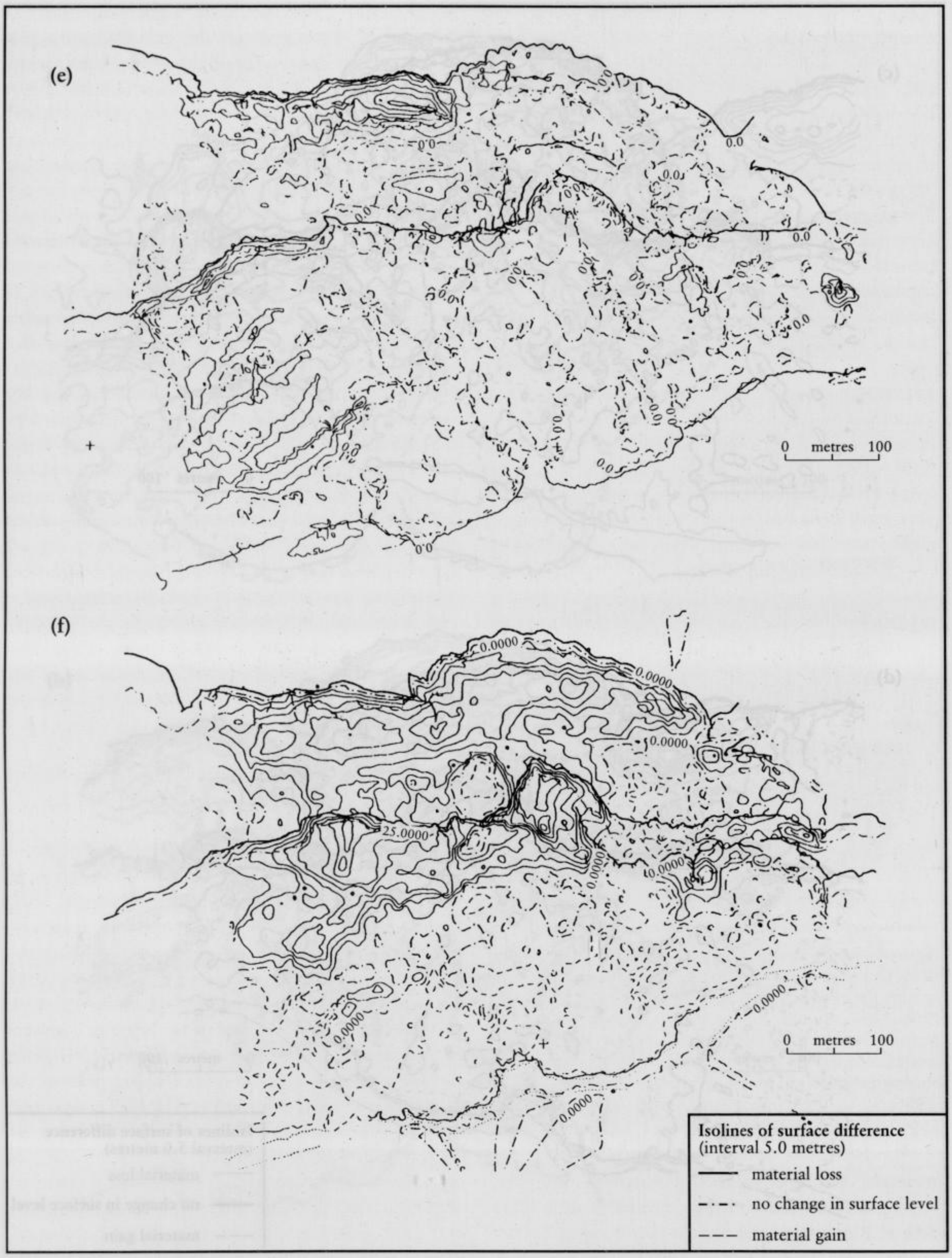


Figure 6.45 – *continued*. Contours of surface difference in elevation (i.e. erosion–deposition–no change) for the periods (e) 1995–1988, and (f) 1988–1946. The 'no change' along the main mudslide axis shows input = output and dynamic equilibrium over a decade interval. After Brunsden and Chandler (1996).

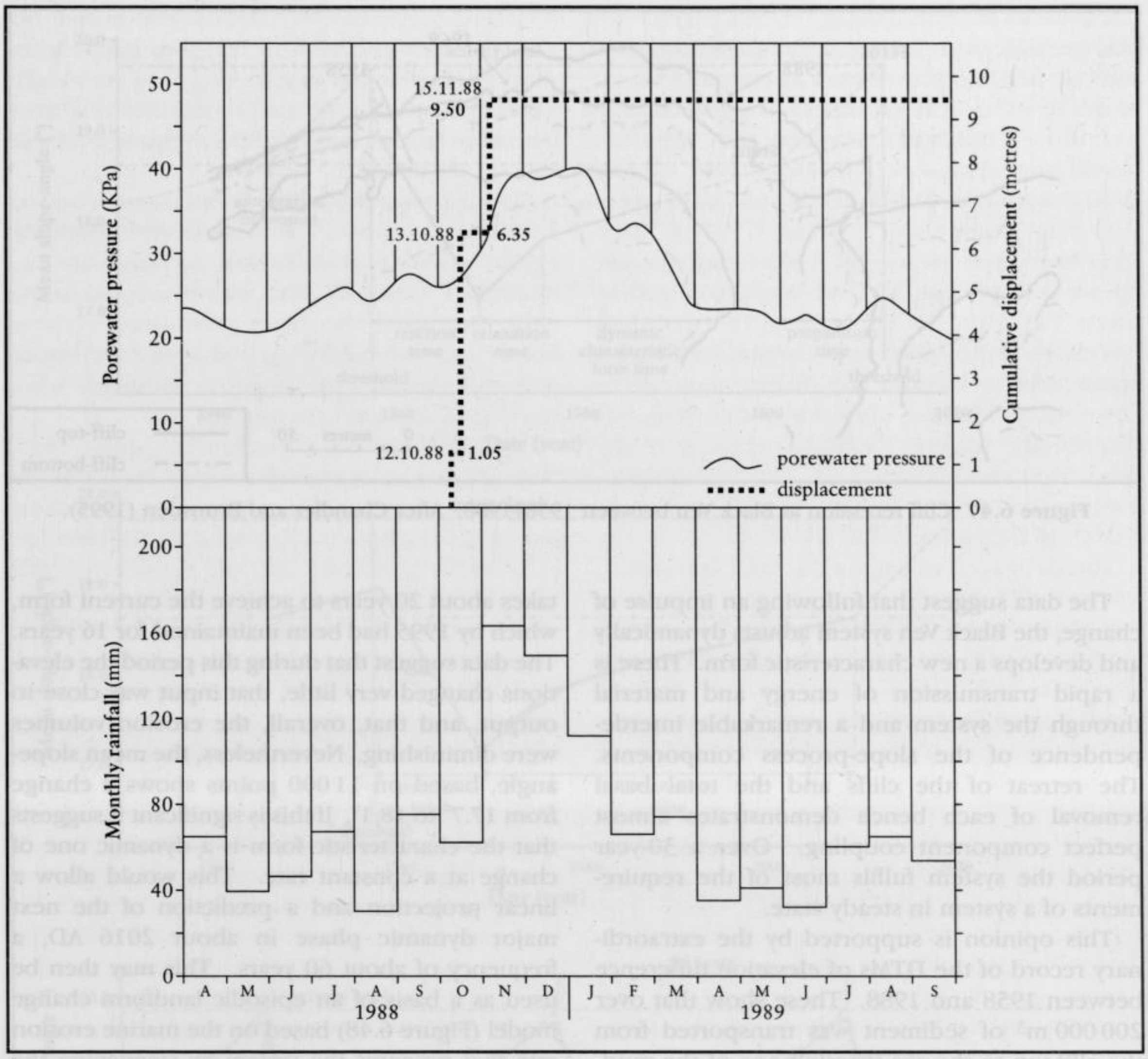


Figure 6.46 The seasonal behaviour of rainfall movement and porewater pressure at Black Ven. The observation that movement ceased on 5/11/88 suggests that there may have been a considerable strength gain following movement. After Koh (1990).

artefact of the fact that data for 1958 are available. The 1958 data therefore represent the first epoch, which happens to have been soon after failure, rather than failure activity at the threshold angle itself.

Following this initial rapid movement, the cliffs adjusted by building lobes of mud into the sea, and in the ensuing ten years the erosional wave diffused upslope to form low-angle slopes on the upper benches. In doing so, the form was maintained even though the whole complex moved inland by as much as 90 m. There was a change in the proportion of low-angle slopes between 1969 and 1988 because the accumulation lobes were being removed by the sea, but the degradation slopes maintained their form. The main processes involved were the cascades of material over the terraces, the parallel retreat of the undercliffs and the rapid transport of material away from the foot of the cliffs and across the benches by the mudslides. Chandler and Brunsden (1995) observe that this is a good example of a retreating but unchanging slope form being maintained by an efficient basal removal condition. Some impression of the rate of recession may be obtained from Figure 6.47.

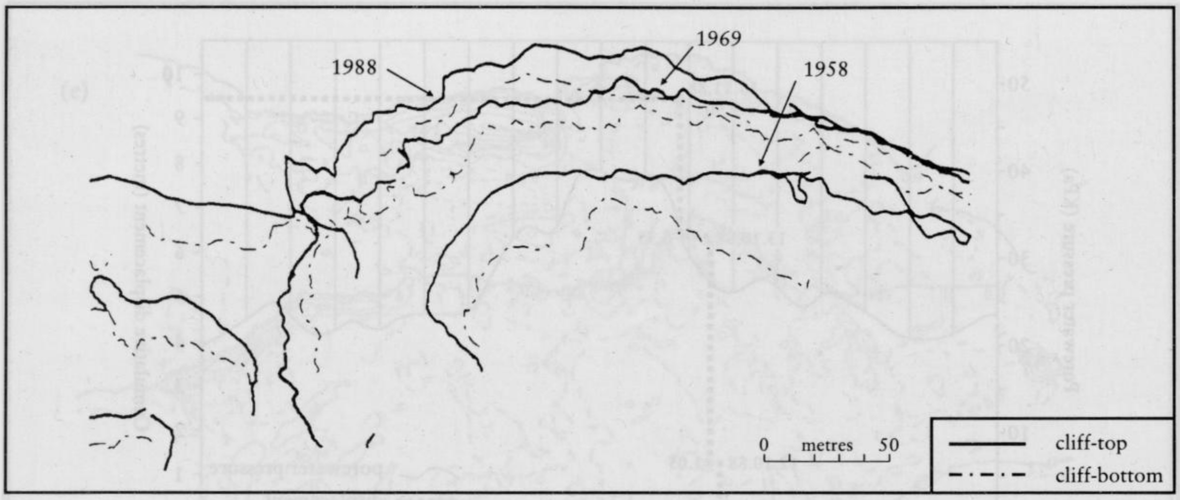


Figure 6.47 Cliff recession at Black Ven between 1958–1988. After Chandler and Brunsden (1995).

The data suggest that following an impulse of change, the Black Ven system adjusts dynamically and develops a new characteristic form. There is a rapid transmission of energy and material through the system and a remarkable interdependence of the slope-process components. The retreat of the cliffs and the total basal removal of each bench demonstrates almost perfect component coupling. Over a 30-year period the system fulfils most of the requirements of a system in steady state.

This opinion is supported by the extraordinary record of the DTMs of elevation difference between 1958 and 1988. These show that over 200 000 m^3 of sediment was transported from the cliff-top to the sea through one of the mudslide systems and yet the overall cliff form remained unchanged to a large extent.

The data may also be used to inform discussion of the timescales of landform change. By manipulating the mean slope-angles at all epochs in a Computer Assisted Design (CAD) system it is possible to produce projections of the mean slope-angle into the future. This can be used to set up a working hypothesis of the possible time adjustments required. The impulse of change or threshold activity crossing is followed by a rapid reduction of slope angles as the mudslides form and the seaward lobes accumulate. This was probably achieved early in the 1960s but the epoch interval (1958–1969) only permits a resolution of the reaction time to 11 years.

The system then relaxed over a further period of 7 years. Therefore, this model suggests that it takes about 20 years to achieve the current form, which by 1995 had been maintained for 16 years. The data suggest that during this period the elevations changed very little, that input was close to output, and that, overall, the erosion volumes were diminishing. Nevertheless, the mean slope-angle, based on 11 000 points shows a change from 17.7° to 18.1°. If this is significant it suggests that the characteristic form is a dynamic one of change at a constant rate. This would allow a linear projection and a prediction of the next major dynamic phase in about 2016 AD, a frequency of about 60 years. This may then be used as a basis of an episodic landform change model (Figure 6.48) based on the marine erosion rate that prepares the system by steepening the slope angle.

However, Chandler and Brunsden (1995) point out that this linear model is almost certainly incorrect. The 1969, 1976 and 1988 data points probably suggest an exponential decay towards the threshold activity angle. In this case, the characteristic form has not yet been achieved, the relaxation time is in progress and a long period of slow slope degradation can be expected. The length of time before the next active phase will then be determined by the rate at which the accumulation lobe, plus any input from upslope, is removed and the cliffs steepened towards the threshold. Chandler and Brunsden (1995) remark that because of the unknown input to the lobes during this basal removal phase, this will no doubt prove to be an example of complex response.

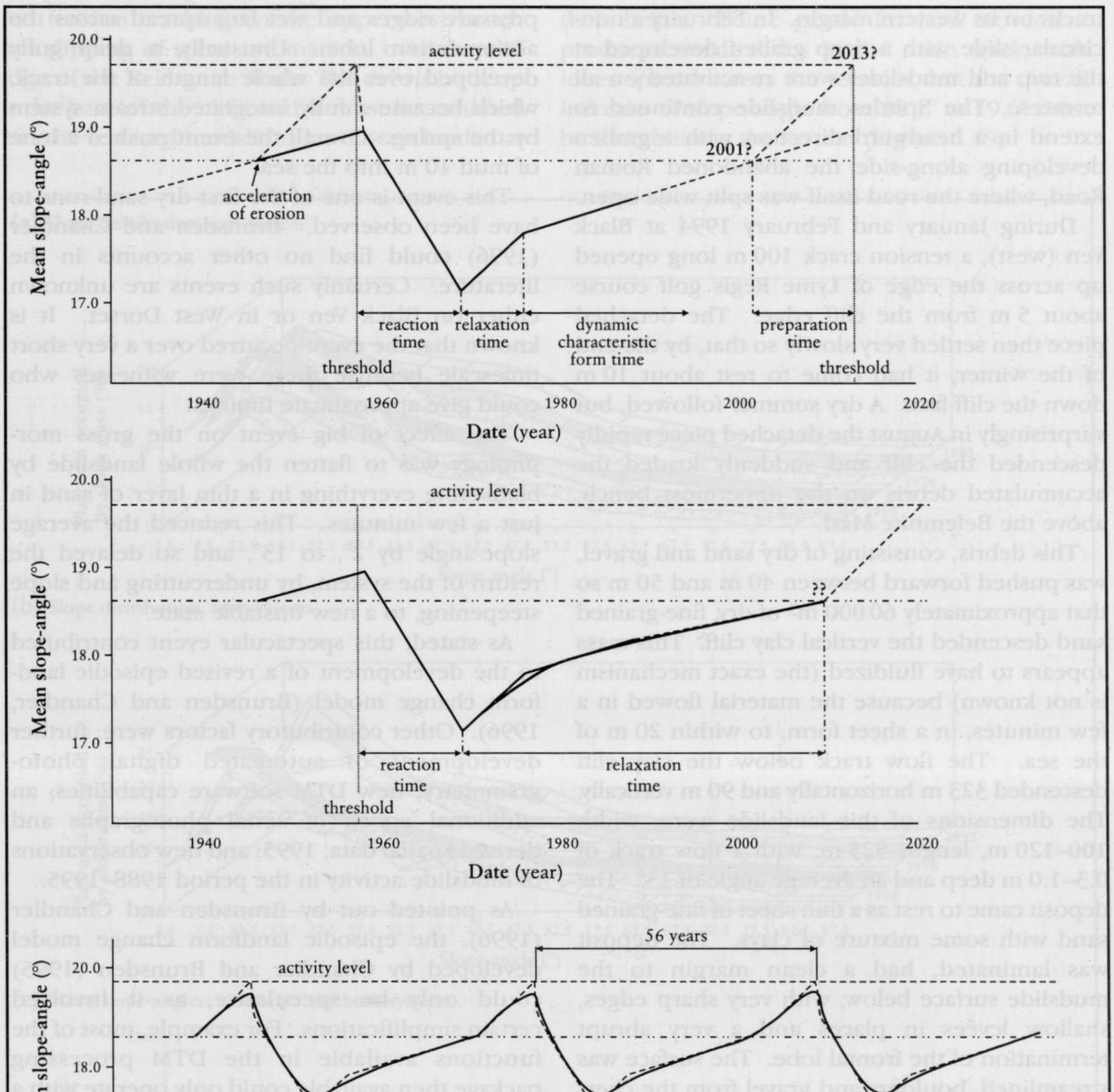


Figure 6.48 Linear model of landslide activity at Black Ven. The lowest diagram is a speculative cyclic model. After Chandler and Brunsden (1995).

(c) The episodic landform change model of Brunsden and Chandler (1996)

A year later, Brunsden and Chandler (1996) substantially revised the 1995 episodic landform change model, partly in response to a dramatic sequence of events that took place on 7 August 1994 at 7.35 in the evening. A large-scale, rare event took place at the western system of mudslides. The previous winter had been one of the wettest on record, with movement observed throughout the cascades. In particular, the dormant systems to the west developed large cracks at their head and very wet failures all along the top of the Belemnite Marl slope. Black Ven (west) began to develop three distinct feeder tracks on its western margin. In February a non-circular slide with a deep graben developed at the toe, and mudslides were re-activated on all terraces. The Spittles mudslide continued to extend in a headward direction with a graben developing along-side the abandoned Roman Road, where the road itself was split wide open.

During January and February 1994 at Black Ven (west), a tension crack 100 m long opened up across the edge of Lyme Regis golf course about 5 m from the cliff edge. The detached piece then settled very slowly so that, by the end of the winter, it had come to rest about 10 m down the cliff-face. A dry summer followed, but surprisingly in August the detached piece rapidly descended the cliff and suddenly loaded the accumulated debris on the uppermost bench, above the Belemnite Marl.

This debris, consisting of dry sand and gravel, was pushed forward between 40 m and 50 m so that approximately 60 000 m^3 of dry, fine-grained sand descended the vertical clay cliff. This mass appears to have fluidized (the exact mechanism is not known) because the material flowed in a few minutes, in a sheet form, to within 20 m of the sea. The flow track below the clay cliff descended 323 m horizontally and 90 m vertically. The dimensions of this landslide were: width 100–120 m, length 525 m, with a flow track of 0.3–1.0 m deep and an average angle of 13°. The deposit came to rest as a thin sheet of fine-grained sand with some mixture of clays. The deposit was laminated, had a clean margin to the mudslide surface below, with very sharp edges, shallow levées in places and a very abrupt termination of the frontal lobe. The surface was streamlined, boulders and gravel from the chert beds were strung out in lines and the overall surface was powdery.

Very shortly after the event all of the mudslides moved forwards, undoubtedly because of the rapid undrained loading of the terraces. On the edge of the Belemnite Marl cliff the loss of the toe of the upper terrace landslides caused a major rotational slide to develop, forming a very prominent scar in the undercliff.

The early autumn of 1994 occasioned significant rainfall. The loose sand, varying in depth between 0.3 m and 1.0 m, quickly became saturated and the surface was transformed into an inaccessible metastable sand. During the very wet winter of 1995 this landslide surface began to sort itself into distinct mud streams with pressure ridges and wet fans spread across the accumulation lobe. Unusually, a deep gully developed over the whole length of the track, which became a fully integrated stream system by the spring. Overall the event pushed a lobe of mud 10 m into the sea.

This event is one of the first dry sand-runs to have been observed. Brunsden and Chandler (1996) could find no other accounts in the literature. Certainly such events are unknown either on Black Ven or in West Dorset. It is known that the event occurred over a very short timescale because there were witnesses who could give approximate timings.

The effect of big event on the gross morphology was to flatten the whole landslide by blanketing everything in a thin layer of sand in just a few minutes. This reduced the average slope-angle by 2°, to 13°, and so delayed the return of the system, by undercutting and slope steepening, to a new unstable state.

As stated, this spectacular event contributed to the development of a revised episodic landform change model (Brunsden and Chandler, 1996). Other contributory factors were: further development of automated digital photogrammetry; new DTM software capabilities; an additional epoch of aerial photographs and derived spatial data, 1995; and new observations of mudslide activity in the period 1988–1995.

As pointed out by Brunsden and Chandler (1996), the episodic landform change model developed by Chandler and Brunsden (1995) could only be speculative, as it involved certain simplifications. For example, most of the functions available in the DTM processing package then available could only operate with a rectangular grid-based DTM. The consequent rectangular and imprecisely specified boundaries to the system resulted in probable distortion to the slope-angle histograms and the mean slope-angles derived from them. Inaccurate specification of boundaries also prevented separate processing and examination of what are in fact two independent mudslide systems. Also, the model was based exclusively on basal erosion; slope steepness triggering of mass-movement activity and other important controlling variables were omitted.

The 1995 epoch shows that the eastern mudslide remained effectively unchanged in form between 1988 and 1995, although very high rainfall in 1994 and 1995 caused some movement and tension cracking on the uppermost bench and slipping of some toe material over the cliff edge of the Belemnite Marl. The debris built up at the base of the cliffs and it must be assumed that the terraced mudslides have been loaded at their heads. Some gentle forward movement produced small slides in the main track, with lobes spread across the terraces, but the overall increase of slope angle from 18.1° to 19.6° must be destabilizing the system (Figure 6.49).

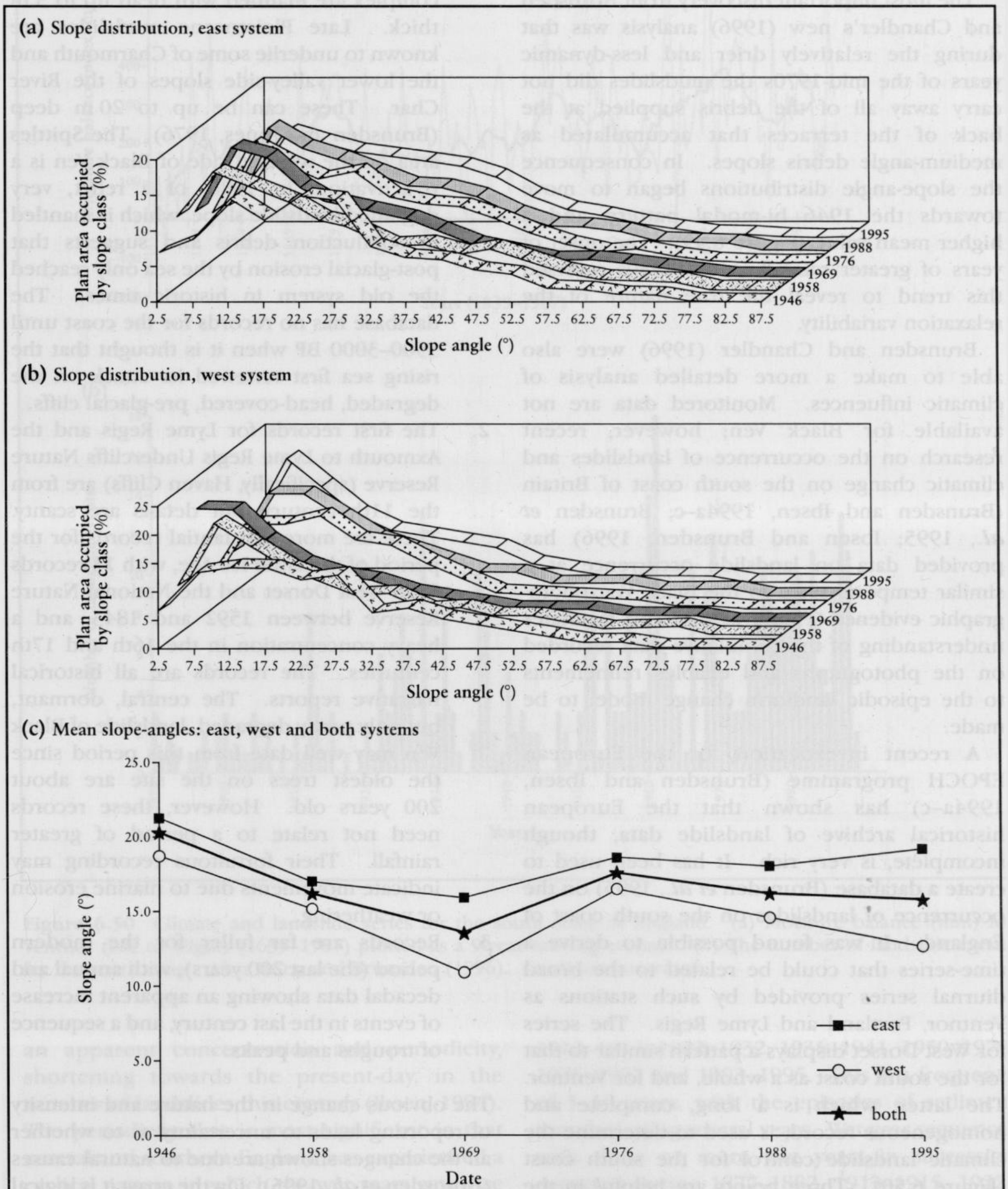


Figure 6.49 Slope-angle graphs for (a) the east system and (b) the west system of Black Ven. After Brunsden and Chandler (1996).

The western landslide continued to show erosion at the head, but with increasing evidence that the system was discharging sediment in a steady-state manner similar to the previous 20 years. There was no change in the morphology and storage of the track.

The most important discovery from Brunsden and Chandler's new (1996) analysis was that during the relatively drier and less-dynamic years of the mid-1970s the mudslides did not carry away all of the debris supplied at the back of the terraces that accumulated as medium-angle debris slopes. In consequence the slope-angle distributions began to move towards the 1946 bi-modal pattern and a higher mean value (Figure 6.49). The onset of years of greater rainfall in the 1980s reversed this trend to reveal the true nature of the relaxation variability.

Brunsden and Chandler (1996) were also able to make a more detailed analysis of climatic influences. Monitored data are not available for Black Ven; however, recent research on the occurrence of landslides and climatic change on the south coast of Britain (Brunsden and Ibsen, 1994a–c; Brunsden *et al.*, 1995; Ibsen and Brunsden, 1996) has provided data on landslide occurrence at a similar temporal scale to the Black Ven photographic evidence (Figure 6.50). This facilitates understanding of the cumulative data recorded on the photographs and enables refinements to the episodic landform change model to be made.

A recent investigation for the European EPOCH programme (Brunsden and Ibsen, 1994a–c) has shown that the European historical archive of landslide data, though incomplete, is very rich. It has been used to create a database (Brunsden *et al.*, 1995) on the occurrence of landslides on the south coast of England. It was found possible to derive a time-series that could be related to the broad diurnal series provided by such stations as Ventnor, Portland and Lyme Regis. The series for West Dorset displays a pattern similar to that for the south coast as a whole, and for Ventnor. The latter, which is a long, complete and homogeneous record, is used to determine the climatic landslide control for the south coast (Figure 6.50). Three points are helpful in the development of the episodic landform change model.

1. The Pleistocene history of Black Ven is not known, but the plateau top, the valley-side slopes to the Lym and Char rivers, the slopes of the neighbouring landslides at Stonebarrow and Golden Cap and the westernmost areas of the Black Ven complex are mantled with head up to 3 m thick. Late Pleistocene mudslides are known to underlie some of Charmouth and the lower valley-side slopes of the River Char. These can be up to 20 m deep (Brunsden and Jones, 1976). The Spittles area on the western side of Black Ven is a re-activation, in 1986, of a relict, very degraded landslide slope, which is mantled in solifluction debris and suggests that post-glacial erosion by the sea only reached the old system in historic times. The database has no records for the coast until 5500–3000 BP when it is thought that the rising sea first renewed its attack on the degraded, head-covered, pre-glacial cliffs.
2. The first records for Lyme Regis and the Axmouth to Lyme Regis Undercliffs Nature Reserve (specifically, Haven Cliffs) are from the 11th century, but details are scanty. There are more substantial records for the period of the Little Ice Age, with 13 records from West Dorset and the National Nature Reserve between 1592 and 1843, and a heavy concentration in the 16th and 17th centuries. The records are all historical narrative reports. The central, dormant, but only partly degraded, landslide of Black Ven may well date from this period since the oldest trees on the site are about 200 years old. However, these records need not relate to a period of greater rainfall. Their fortuitous recording may indicate movements due to marine erosion or weathering.
3. Records are far fuller for the modern period (the last 200 years), with annual and decadal data showing an apparent increase of events in the last century, and a sequence of troughs and peaks.

The obvious change in the nature and intensity of reporting leads to uncertainty as to whether all the changes shown are due to natural causes (Brunsden *et al.*, 1995). On the coast it is logical to ascribe some of the increase to sea-level rise and erosion of the sea cliffs. There appears to be

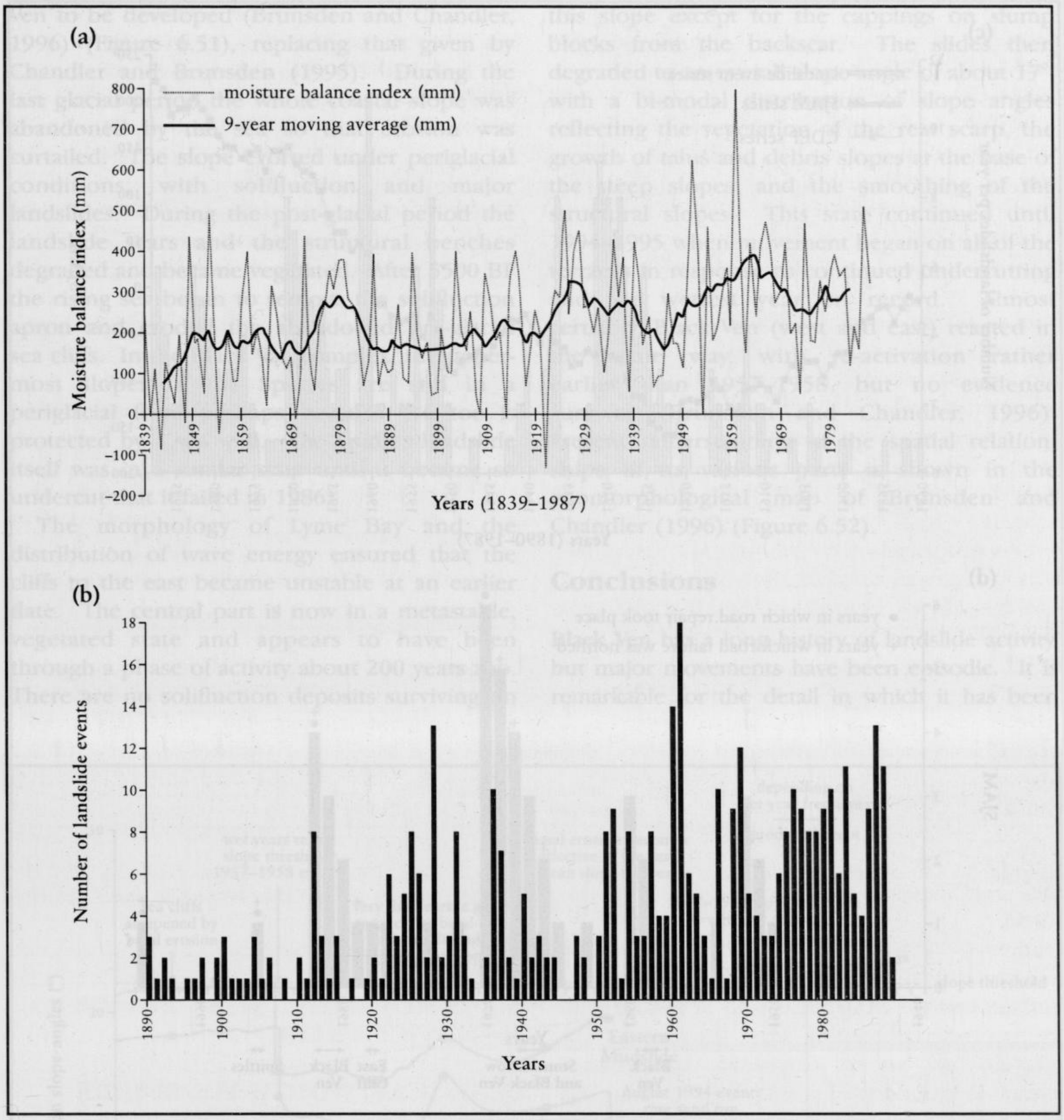


Figure 6.50 Climate and landslide series for the south coast of England. (a) Moisture balance (mm) for Ventnor (Isle of Wight) (1639–1987) plotted as a 9-year moving average. (b) The number of landslide events for the south coast. After Ibsen and Brunsden (1996). *Continued overleaf.*

an apparent concentration and periodicity, shortening towards the present-day, in the number of landslides this century (Ibsen, 1994). The periodicity that is most significant in that central and southern England has experienced a cyclical pattern in which the wet years gave rise to greater geomorphological activity. There is a concentration of landsliding during the periods 1912–1913, 1922–1932, 1936–1941, 1950–1970, 1975–1982 and 1993–1995. This is a frequency of 5–10 years, with the episodes of sediment transfer lasting several years. Wet-year sequences with three or more wet years in succession occurred during 1877–1882, 1913–1915, 1922–1932, 1936–1939, 1952–1954, 1963–1970 and 1978–1982, all co-inciding with landslide

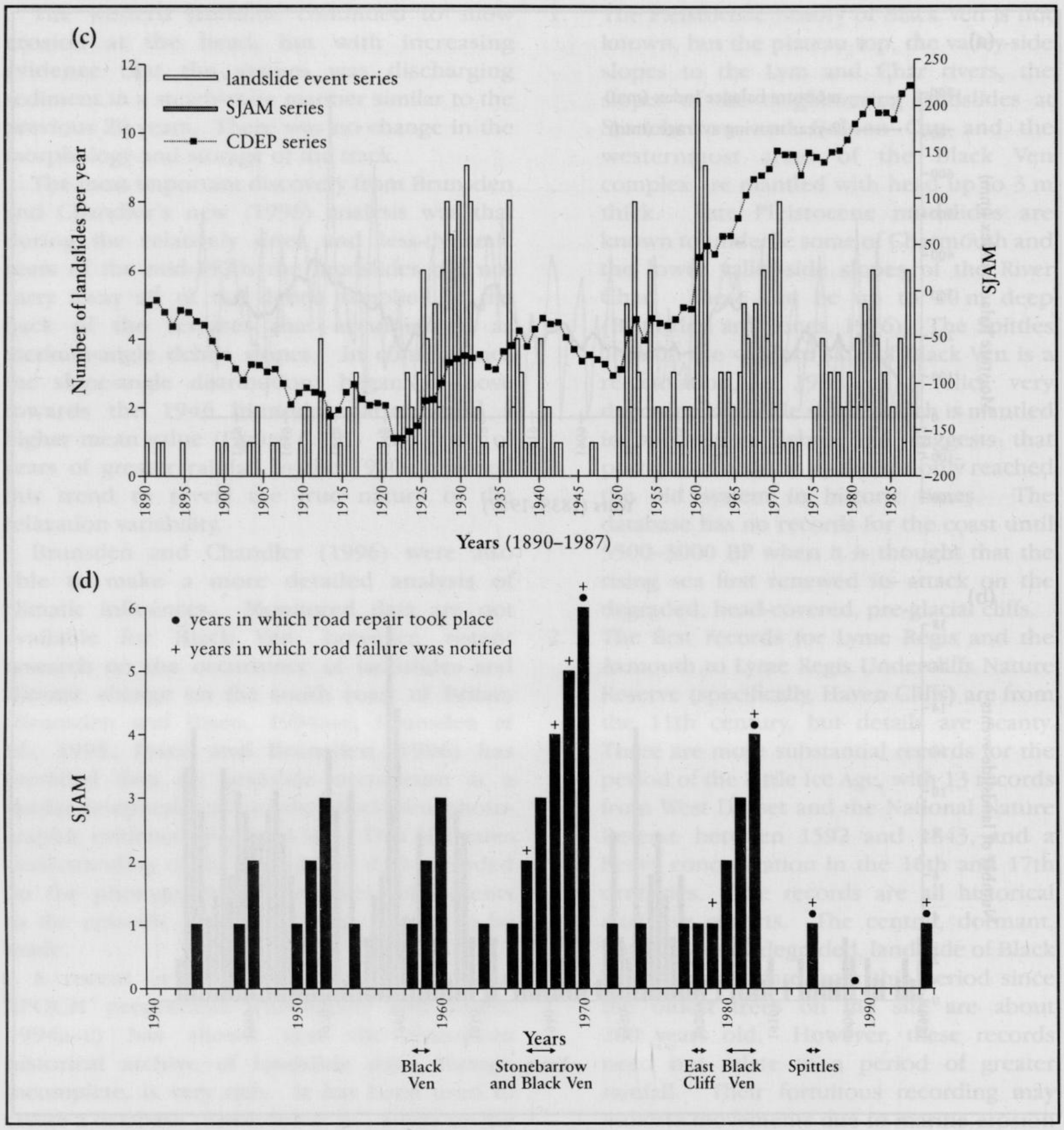


Figure 6.50 – *continued*. Climate and landslide series for the south coast of England. (c) The cumulative moisture balance departure from the mean (CDEP), the cumulative number of years with moisture balance greater than the mean (SJAM) and the landslide occurrence at Ventnor (after Ibsen and Brunsden, 1996). (d) The sequence of years of higher rainfall and landslide occurrences for west Dorset (after Brunsden and Chandler, 1996).

records. The precipitation record for the Axmouth to Lyme Regis Undercliffs National Nature Reserve broadly supports this pattern.

It therefore seems reasonable to accept, for the modelling of the episodic behaviour of the West Dorset landslides, a 'wet' year climatic control of 5–10 years lasting for 3–6 years which is superimposed on the trend for increasing wetness, sea-level rise, and slope steepening over the last 60 years.

These records allow a further speculative model for episodic landform change at Black

Ven to be developed (Brunsden and Chandler, 1996) (Figure 6.51), replacing that given by Chandler and Brunsden (1995). During the last glacial period the whole coastal slope was abandoned by the sea so that erosion was curtailed. The slope evolved under periglacial conditions, with solifluction and major landslides. During the post-glacial period the landslide scars and the structural benches degraded and became vegetated. After 5500 BP the rising sea began to remove the solifluction apron and eroded the abandoned pre-glacial sea cliffs. In the Black Ven complex, the uppermost slopes of the Spittles are still in a periglacial form perhaps because the toe is protected by a sea wall. The Spittles landslide itself was in a similar state until it became so undercut that it failed in 1986.

The morphology of Lyme Bay and the distribution of wave energy ensured that the cliffs to the east became unstable at an earlier date. The central part is now in a metastable, vegetated state and appears to have been through a phase of activity about 200 years ago. There are no solifluction deposits surviving on this slope except for the cappings on slump blocks from the backscar. The slides then degraded to an overall slope-angle of about 17°, with a bi-modal distribution of slope angles reflecting the vegetation of the rear scarp, the growth of talus and debris slopes at the base of the steep slopes, and the smoothing of the structural slopes. This state continued until 1994–1995 when movement began on all of the terraces in response to continued undercutting and the wettest year on record. Almost certainly, Black Ven (west and east) reacted in the same way, with re-activation rather earlier than 1957–1958, but no evidence survives (Brunsden and Chandler, 1996). Present understanding of the spatial relationships of its various parts is shown in the geomorphological map of Brunsden and Chandler (1996) (Figure 6.52).

Conclusions

Black Ven has a long history of landslide activity, but major movements have been episodic. It is remarkable for the detail in which it has been

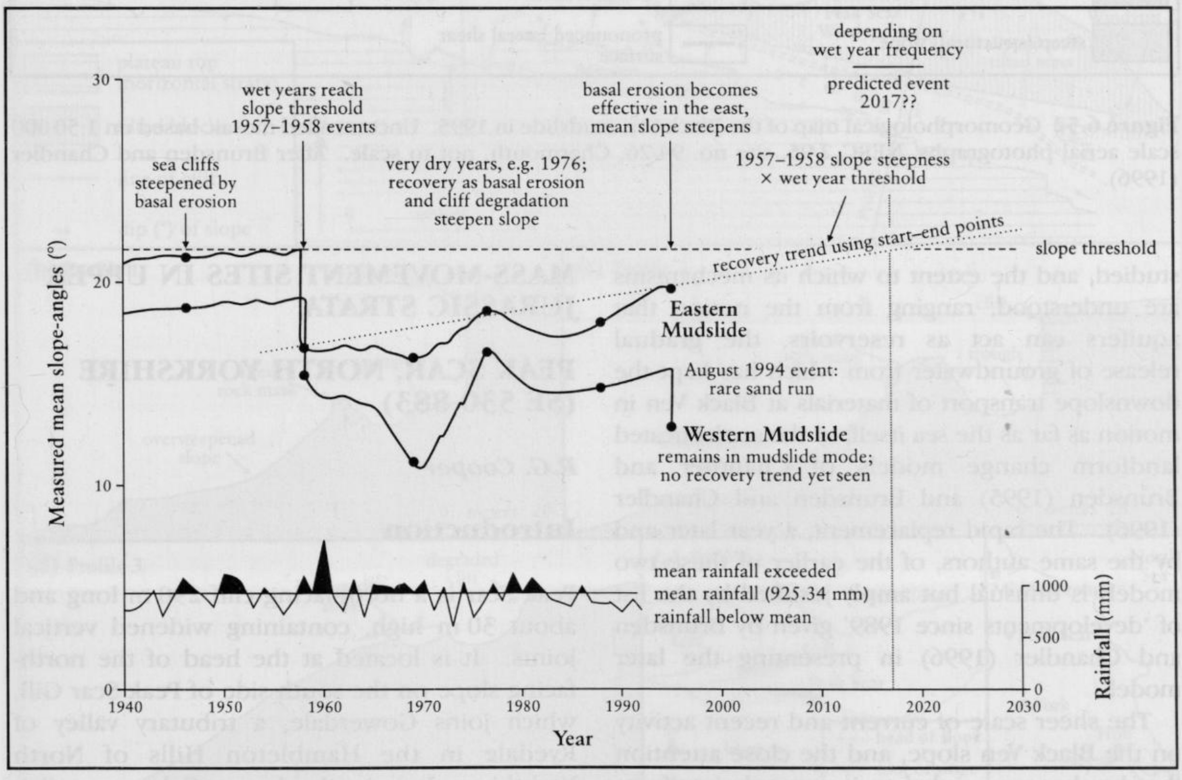


Figure 6.51 A temporal model of episodic landform change at Black Ven. After Brunsden and Chandler (1996).

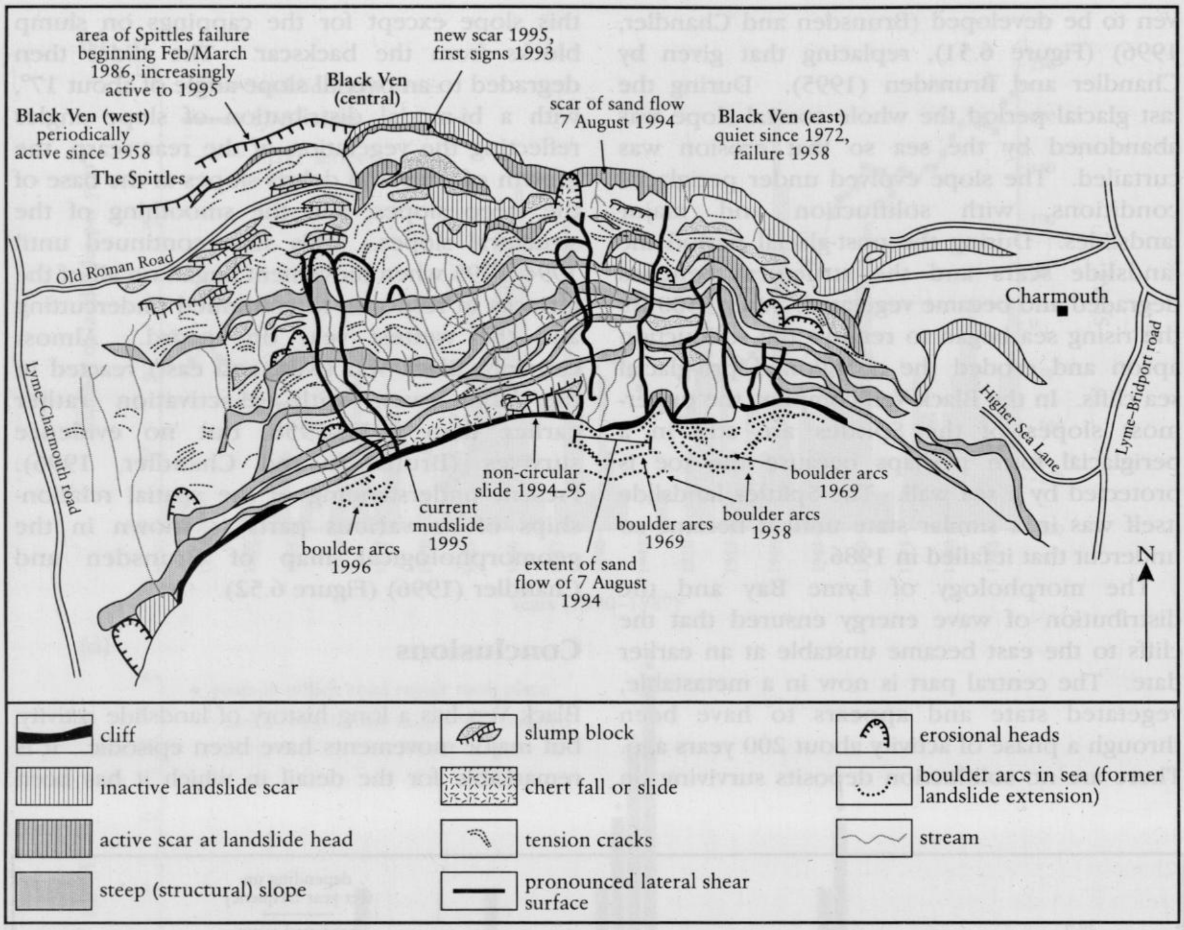


Figure 6.52 Geomorphological map of the Black Ven mudslide in 1995. Uncontrolled mosaic based on 1:50 000 scale aerial photographs, NERC 2/95, site no. 94/26, Charmouth, not to scale. After Brunsden and Chandler (1996).

studied, and the extent to which its mechanisms are understood, ranging from the notion that aquifers can act as reservoirs, the gradual release of groundwater from which has kept the downslope transport of materials at Black Ven in motion as far as the sea itself, to the sophisticated landform change models of Chandler and Brunsden (1995) and Brunsden and Chandler (1996). The rapid replacement, a year later and by the same authors, of the earlier of these two models is unusual but amply justified by the list of 'developments since 1989' given by Brunsden and Chandler (1996) in presenting the later model.

The sheer scale of current and recent activity on the Black Ven slope, and the close attention that has been accorded to it, strongly justify its inclusion as a Geological Conservation Review site.

MASS-MOVEMENT SITES IN UPPER JURASSIC STRATA

PEAK SCAR, NORTH YORKSHIRE (SE 530 883)

R.G. Cooper

Introduction

Peak Scar is a north-facing cliff 290 m long and about 30 m high, containing widened vertical joints. It is located at the head of the north-facing slope on the south side of Peak Scar Gill, which joins Gowerdale, a tributary valley of Ryedale in the Hambleton Hills of North Yorkshire. It is in the Lower Calcareous Grit Member of the Corallian (Upper Jurassic) strata. Between 9 m and 21 m downslope from the foot of the cliff is a row of massive detached blocks of rock, tilting away from the cliff and towards the valley, with an apparent valley-ward dip of 26°. The main mass is 130 m long and nearly 30 m broad. The crest of the ridge formed by these detached blocks rises 15 m above the floor of the trough that separates it from the cliff-face. The slope on the downslope side of the detached blocks is steeper than adjacent slopes to the west and east of Peak Scar (Figure 6.53).

Description

The sequence of features, on moving downslope, at Peak Scar is (Cooper, 1982):

Vegetated plateau top/crest of slope
Vertical cliff-face in bedrock
Poorly vegetated trough or trench with large boulders; upper sides of detached blocks, in bare or moss-clad rock with pronounced downslope tilt
Vegetated top and downslope sides of detached blocks
Undulating slope to stream course at the slope foot

This sequence is remarkably similar to one of the sites described by de Freitas and Watters (1973) in their landmark paper on toppling failure. The site in Nant Gareg-lwyd, in the Rhondda Valley near Blaenrhondda, Powys, shows a joint-controlled vertical cliff-face in hard rock, at the foot of which lies a large toppled mass dipping downslope, and separated from it by a hillside trench. Peak Scar displays all of these features far more clearly, and with greater vertical development (see Figure 6.53).

The Corallian strata are a predominantly horizontally bedded, alternating series of calcareous sandstones and oolitic limestones (the Hambleton Oolite). The areal distribution of two kinds of features on the slopes of the Hambleton Hills suggest that a particular type of landsliding is taking place in this area. Firstly, enterable widened joints, locally termed 'windypits' are found on the slopes and on the surface of the plateau, close to its edges (Fitton and Mitchell, 1950; Cooper *et al.*, 1976). Secondly, ridge-and-trough features (Briggs and

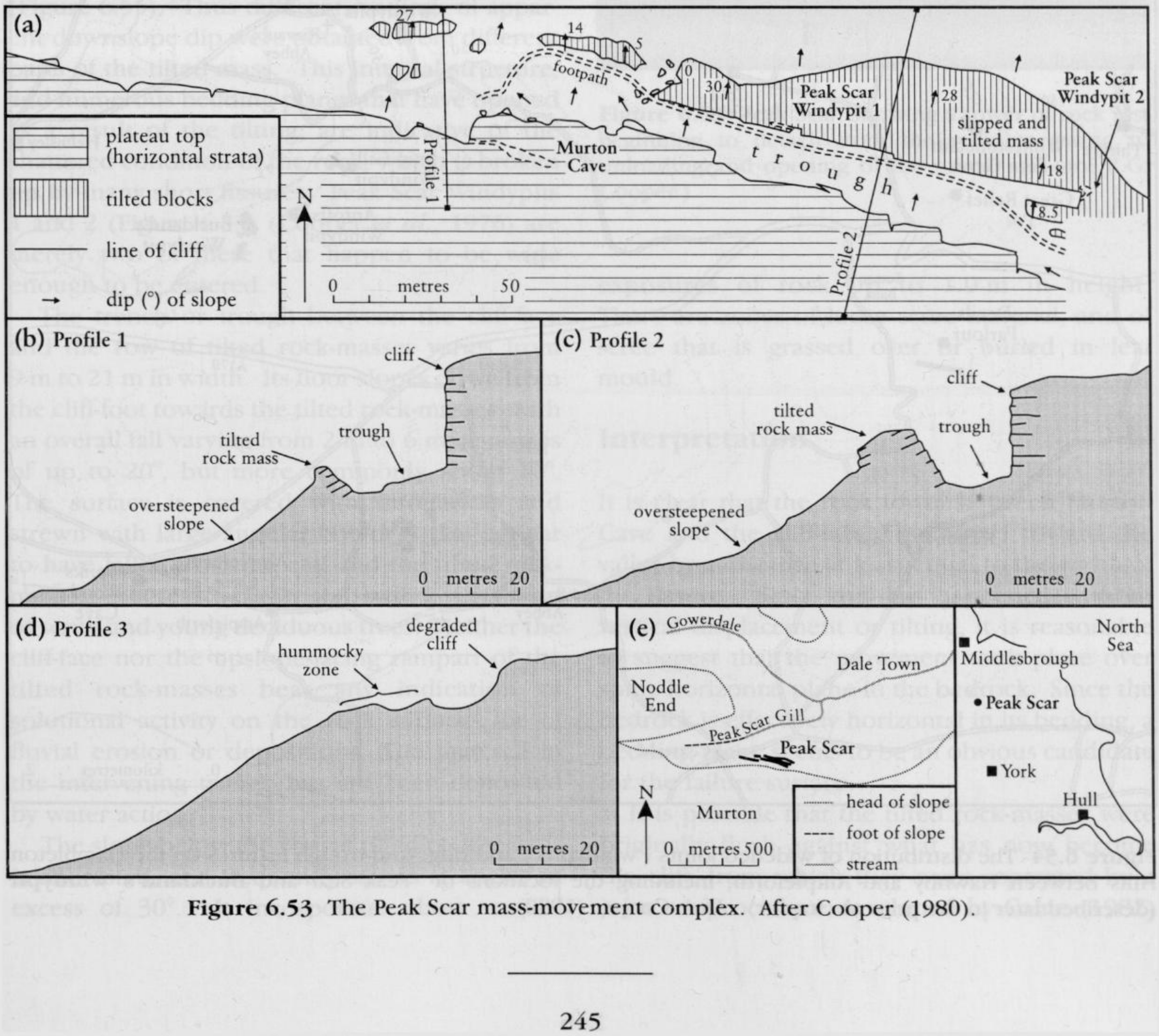


Figure 6.53 The Peak Scar mass-movement complex. After Cooper (1980).

Courtney, 1972), hillside trenches and uphill-facing scarps (Radbruch-Hall, 1978) are found on the slopes. Sites of enterable widened joints, and of ridge-and-trough features are shown in Figure 6.54.

At Peak Scar, the cliff-face is intersected by joints running sub-normal to it, which offset its line towards the valley at several places. These joints, and a series of pronounced horizontal bedding planes, divide the cliff-face into large

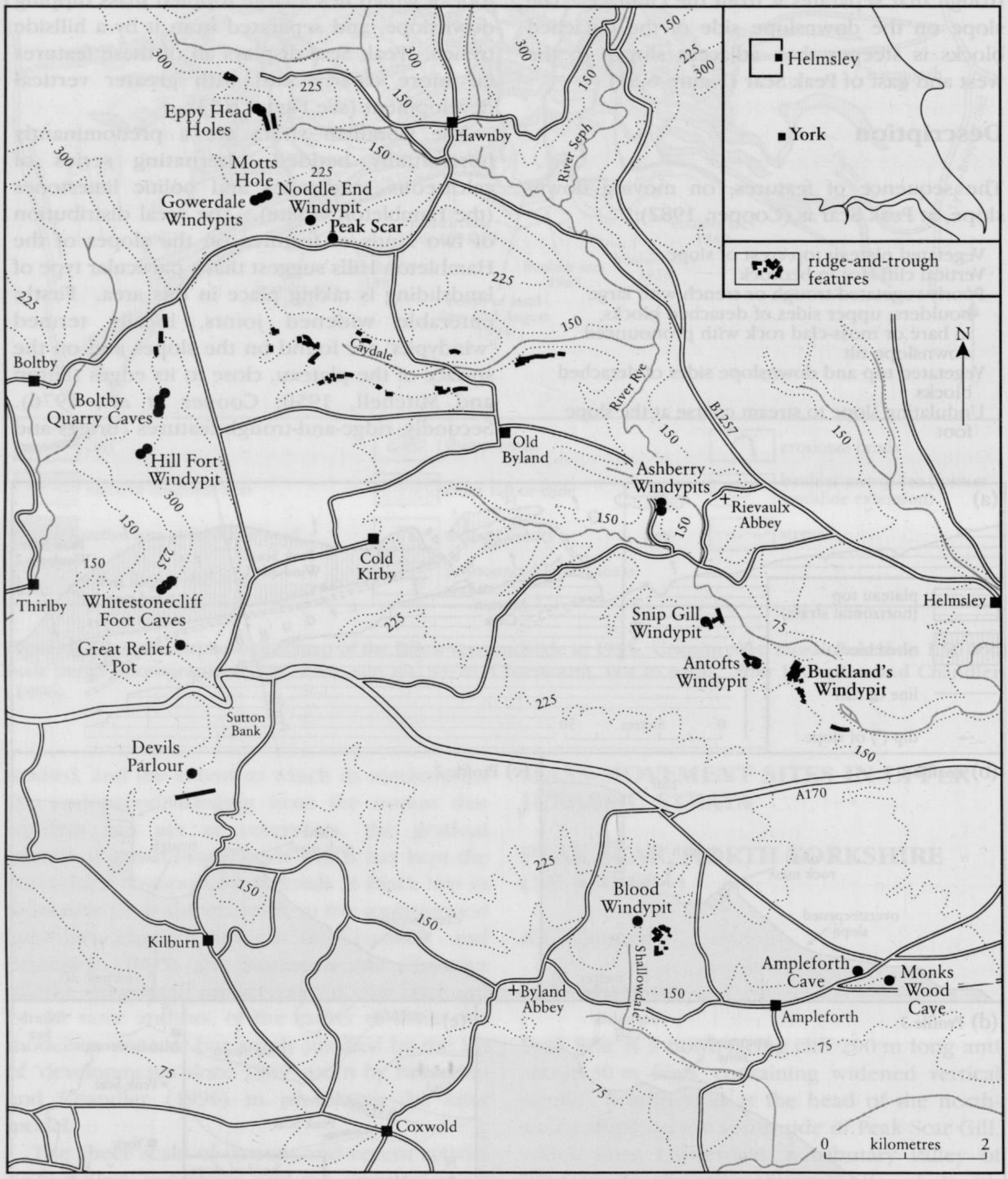


Figure 6.54 The distribution of widened joints ('windypits') and ridge-and-trough features on the Hambleton Hills between Hawnby and Ampleforth, including the locations of Peak Scar and **Buckland's Windypit** (described later in the present chapter). After Cooper (1980).

sub-cuboidal blocks. Many of the blocks appear to be loose as they are separated from the main mass of the rock by joints that run sub-parallel to the cliff-face, but behind it. About 70 m from its eastern end the line of the cliff-face is offset 8 m away from the valley by a joint. At the point of the inner corner produced by this offset is the entrance to Murton Cave, which is an enterable fissure running sub-parallel to the cliff-face.

Murton Cave (Cooper *et al.*, 1976) has been formed by the widening of a joint. Its upslope wall is continuous with the set-back face of the cliff. Between the fissure and the cliff-face in front of it (i.e. on its valley-ward side) is a tower of Corallian limestone 8 m thick and 30 m high, the upslope side of which forms the downslope wall of the fissure. The downslope wall of Murton Cave does not appear to have been displaced upwards or downwards, and appears to have undergone no tilting with respect to the upslope wall.

Some of the tilted rock-masses seem to be made up of several towers of rock, each tilted and leaning against the next one downslope (Figure 6.55). Thus different readings of apparent downslope dip were obtained from different parts of the tilted mass. This internal structure, and numerous bedding planes that have opened as a result of the tilting, are indicative of the shattered condition of the rock, which is broken up by many short fissures. Peak Scar Windypits 1 and 2 (Figure 6.53) (Cooper *et al.*, 1976) are merely two of these that happen to be wide enough to be entered.

The trench or trough between the cliff-face and the row of tilted rock-masses varies from 9 m to 21 m in width. Its floor slopes down from the cliff-foot towards the tilted rock-masses, with an overall fall varying from 2 m to 6 m, at angles of up to 20°, but more commonly about 10°. The surface is covered with hummocks and strewn with large angular boulders that appear to have fallen from the cliff and the tilted rock-masses. There is a clayey soil, with a vegetation of scrub and young deciduous trees. Neither the cliff-face nor the upslope-facing rampart of the tilted rock-masses bear any indication of solutional activity on the rock surfaces, or of fluvial erosion or deposition. The material in the intervening trough has not been deposited by water action.

The slope below the row of tilted rock-masses is afforested, and frequently stands at angles in excess of 30°. It incorporates short vertical

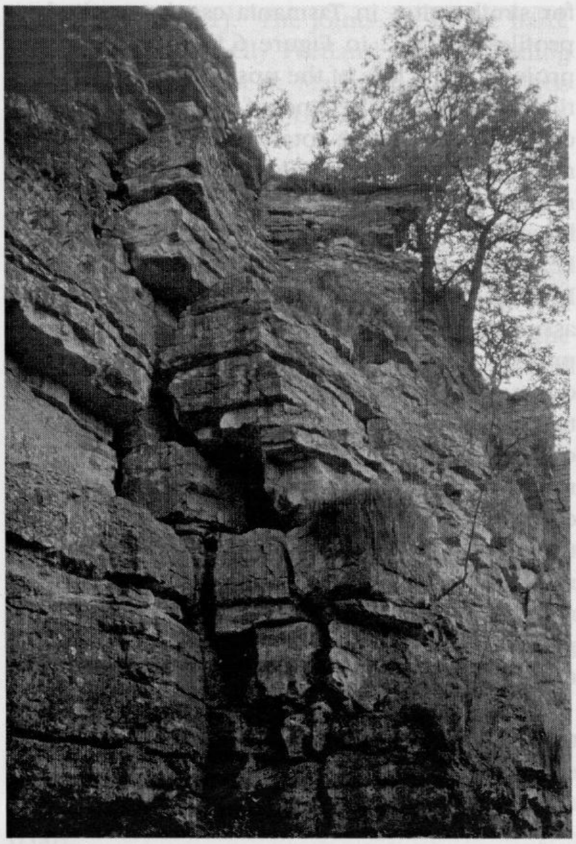


Figure 6.55 Peak Scar, showing a tower of rock just beginning to detach from the rockface owing to unloading and opening of the joints. (Photo: R.G. Cooper.)

exposures of rock up to 1.0 m in height. There are zones of loose scree material, and of scree that is grassed over or buried in leaf mould.

Interpretation

It is clear that the rock tower between Murton Cave and the cliff-face has moved toward the valley by a distance at least equal to the width of the fissure. Since this has happened without vertical displacement or tilting, it is reasonable to suggest that the movement took place over some horizontal plane in the bedrock. Since the bedrock is effectively horizontal in its bedding, a bedding plane seems to be an obvious candidate for the failure surface.

It is possible that the tilted rock-masses were originally flush against what has now become the cliff, from which they were separated by a joint. A procedure developed by Caine (1982) for similar sites in Tasmania can be applied to profiles 1 and 2 in Figure 6.53. This involves projecting the line of the upslope rampart of the tilted mass and the line of the cliff until they cross. At this point rotation of the tilted mass may have taken place (Figure 6.56). It is worth noting that Caine's sites are about three times larger (in each dimension) than the features at Peak Scar, but the same considerations apply. Reconstruction in the manner of Caine (1982), assuming the later movement to have been purely rotational (Cooper, 1980), indicates a downslope rotation of 33° about a line parallel to and in line with the cliff, 40 m below the top of the cliff and 22 m below the trench floor. On reconstruction in this way, it is apparent that a considerable amount of rock has been lost from the tilted masses, as when set upright they do not reach the cliff-top. It is suggested that this material has slid off downslope over bedding planes, and is in part responsible for the steepness of the lower slope.

It has been suggested (Cooper, 1982) that three major types of mechanism acting in succession may be adduced as an explanation of the suite of features at Peak Scar.

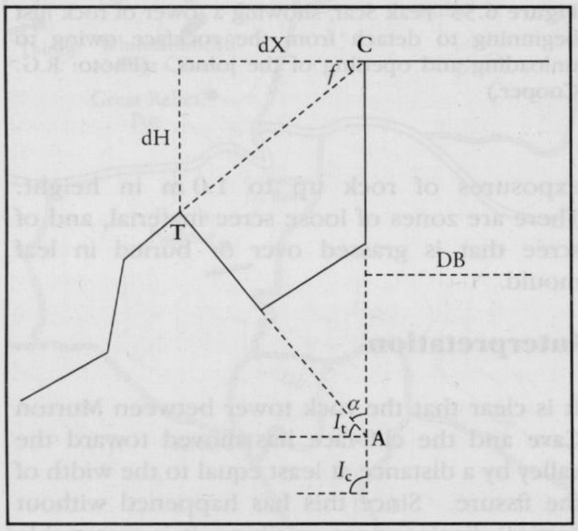


Figure 6.56 Terms and definitions used in modelling a hypothetical topple: (A: point of convergence based on extrapolation from joint surveys on the topple and the cliffs; C: cliff crest; T: crest of the topple; DB: elevation at the base of the dolerite sheet; LC: distance a–c; LT: distance a–t; I_c: joint inclination on the cliff; I_t: joint inclination on the topple; α: tilt angle of the topple (I_c–I_t); dH: vertical lowering of the topple crest; dX: lateral displacement of the topple crest; *f*: angle from c–t). After Caine (1982).

1. *Splitting mechanisms*: clearly, fissures must form in a massive rock before they can be widened. They could be tectonically initiated joints, and thus pre-date not only the slope features, but also the incision of the valley itself. Alternatively they could be 'valley joints', formed as a response to stress-release on valley incision.
2. *Sliding mechanisms*: as the initial translational movement seems to be effectively horizontal, it is likely that the plane of sliding is horizontal. It could be located at the junction with the Oxford Clay at the base of the Corallian. However, there is a broad lithological transition zone between the two. Alternatively, there could be thin 'mobile' bands of clay or shale interbedded in the Corallian strata, which would provide planes of sliding.
3. *Tilting mechanisms*: the tilting towards the valley of blocks that have slid forward could be due to undermining of blocks by erosional removal of weaker material beneath, and the settlement of the blocks. Alternatively, it is possible that the blocks move forward until their centres of gravity are unsupported due to the steepness of the slope, and then over-balance. Either way, the tilting seems to represent a form of toppling failure (de Freitas and Watters, 1973). The observation of uphill-facing scarps describes features widely regarded as being diagnostic of 'sagging failures'.

Conclusions

Peak Scar is the best British example of a de Freitas and Watters' intermediate-sized toppling failure, by virtue of its apparent 'freshness' (which may be deceptive) and its vertical range. The characteristic of the downslope wall of Murton Cave, that does not appear to have been displaced upwards or downwards, and has not rotated with respect to the upslope wall, is common to most of the other widened joints in the Hambleton Hills. The widespread distribution of both widened joints and hillside trenches in the Hambleton Hills suggests that the three-fold process evident at Peak Scar may have wider applicability than just the immediate site. Other interesting sites include the East Weare and Great Southwell slips on the Isle of Portland, Dorset, and Daddyshole Plain and Ansteys Cove, Torquay, Devon.

BLACKNOR CLIFFS, ISLE OF PORTLAND, DORSET (SY 677 715)

R.G. Cooper

Introduction

The cliffs on the western coast of the Isle of Portland (Figures 6.57 and 6.58) exhibit probably the best British examples of toppling failures. This type of failure involves the separation of a monolith or slab of hard rock from the surrounding rock, usually by jointing. At a cliff edge, the monolith or slab will be isolated between the cliff-face and the joint(s). Eventually the slab will fall. This site shows various stages of the process.

Toppling failure, which involves forward rotation of the mass, is classically preceded by a period of accelerating creep, which may last for several years. This may involve slow outward movement of the slab away from the surrounding rock, widening the joints, over some underlying incompetent horizon (Schumm and Chorley, 1964).

Portland is traversed by a series of NNE–SSW-trending major joints. These have been mapped by Coombe (1981). The joints are in the hard, often oolitic or shelly Portland Stone, running upwards into Purbeck Marls and downwards into Portland Sands (all Upper Jurassic).

Description

There are three distinct lines of evidence that this type of failure is occurring on the Isle of Portland. Firstly, the NNE–SSW joints are almost all widened to a greater or lesser degree. Some of them in the interior of Portland have been entered and explored, via the quarries on its plateau surface (Ford and Hooper, 1964; Churcher *et al.*, 1970; O'Conner and Graham, 1996). It is likely that the joints owe both their existence and their widening to Alpine Earth movements (Cooper, 1983a,b). Around the coast of Portland, the joints are wider than in the interior, presumably due to seaward movement of the huge slabs of limestone isolated between them and the cliff-line. On the western side of Portland, between Blacknor and Mutton Cove, these joints have been entered and extensively mapped (Cooper and Solman, 1983; Graham and Ryder, 1983; Cooper *et al.*, 1995) (Figures 6.59 and 6.60).

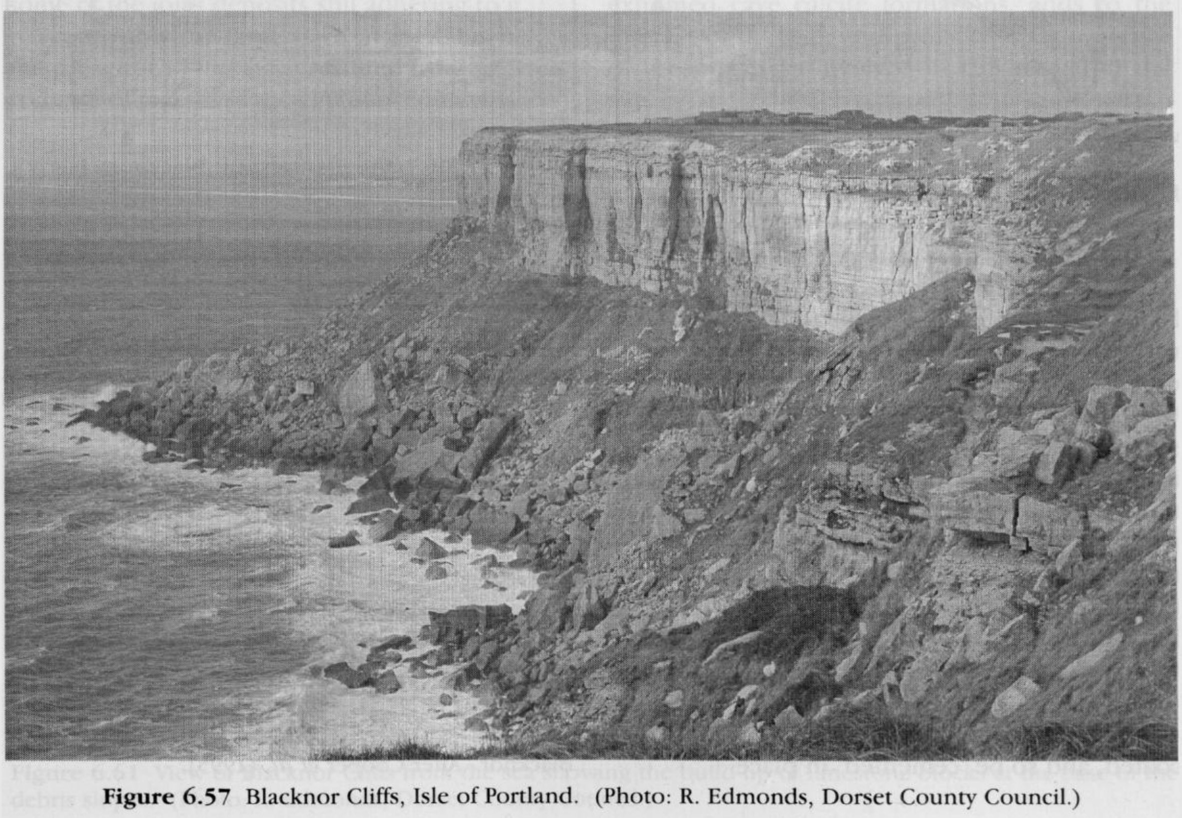


Figure 6.57 Blacknor Cliffs, Isle of Portland. (Photo: R. Edmonds, Dorset County Council.)

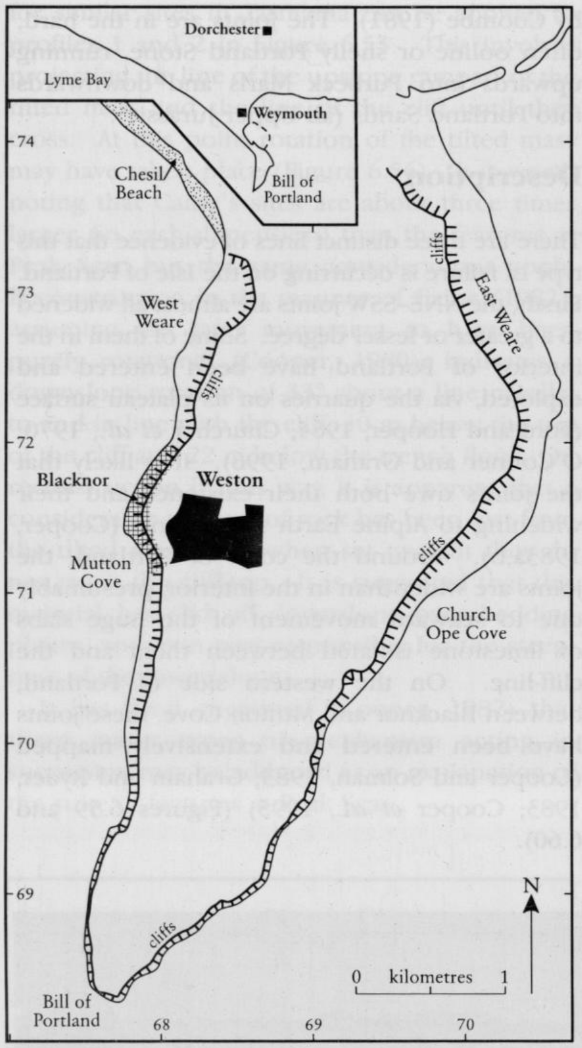


Figure 6.58 Location of the Blacknor Cliffs GCR site.

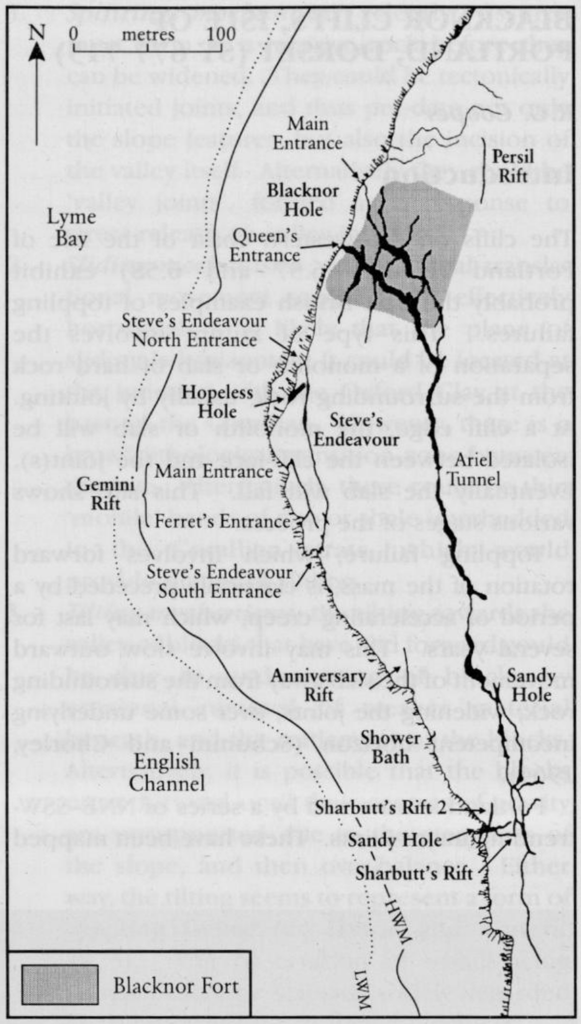


Figure 6.59 The fissures of Blacknor, Isle of Portland.

Secondly, at Blacknor, the 30 m-high cliff-face bears speleothems, principally calcite flowstone (travertine) coating the limestone faces, and often developed into 'organ-pipe' formations. Unsorted assemblages of clasts ranging from sands to boulders in size are cemented to the cliff-face by this calcite, at various heights above the cliff-foot (Cooper and Solman, 1983). It is difficult to imagine how a boulder, falling over the cliff, could be arrested part-way down by a thin film of deposited calcite on the cliff-face. However, it is easy to imagine how a boulder, falling into an open fisuure, could become wedged part-way down and then coated with calcite at the same time that the walls become coated, and so be 'cemented' in place.

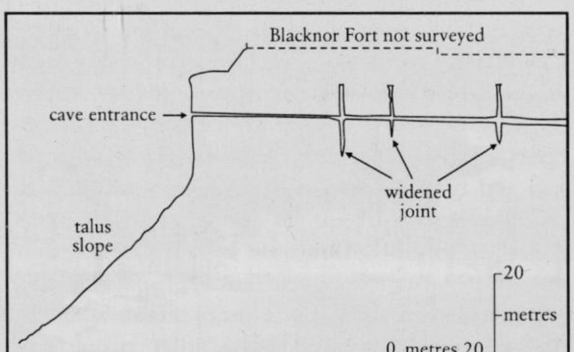


Figure 6.60 Cross-section through the cliff at Blacknor. After Cooper *et al.* (1995).

Thirdly, the 120 m-long debris slope below the cliffs, standing at angles of up to 40°, is littered at its lower end with large, joint-faced limestone blocks, chiefly Lower Purbeck and Portland Stone (Figure 6.61). Some of these bear patches of calcite flows. The blocks are up to 42 m long and 14 m broad, but the third dimension tends to be only a few metres, giving a marked tabular 'slab' shape. Individual units are up to 3500 m^3 in volume.

Interpretation

The wideness of the joints, particularly those near the cliff edge, is a clear indication that the blocks isolated by them have moved laterally, causing the present cliff-line itself to move seawards. Presumably the movement has taken place by bedding-plane shearing on an incompetent horizon in the 'parallel bands' of the Portland Sand, or even in the underlying Kimmeridge Clay.

The speleothem calcite deposits on subaerial cliff-faces suggest that major lengths of the present cliff-face were once the landward walls of widened joints, and that the rock masses which were once on the seaward sides of these joints have collapsed, leaving the present face with some of the joint deposits still adhering to it.

Presumably the large slabs in the talus below the present cliff-line have resulted from periodic collapse of masses of the cliff between the (then) cliff-face and joints parallel to and behind it (Cooper and Solman, 1983). It may be expected that the slabs currently isolated between the present cliff-face and the enterable widened joints behind it will eventually collapse.

Brunsden *et al.* (1996b) associate the joint widening with a large-scale process which accounts for the 2 km-long physiographic 'low' in part of the interior of Portland. Applying to Portland the ideas developed by Cancelli and Pellegrini (1987) in a study of the Northern Appenines in Italy, they propose differential settling of the 'prisms' of rock separated from each other by the major joints. This model 'has not been tested by sub-surface exploration' and needs to be checked against the observable features within the fissures (Brunsden, 1996b).

Conclusions

The site illustrates the sequence of events more clearly than any other site of toppling failure in Great Britain. The joint control of the process is very apparent. The intact collapsed slabs are much larger than those found elsewhere. The fortuitous preservation of unsorted debris up to boulder size, cemented to the once underground (but now subaerial) cliff-face by exhumed cave calcite formations, adds to the interest of the site, and provides clear evidence of the processes involved in cliff retreat at the site.

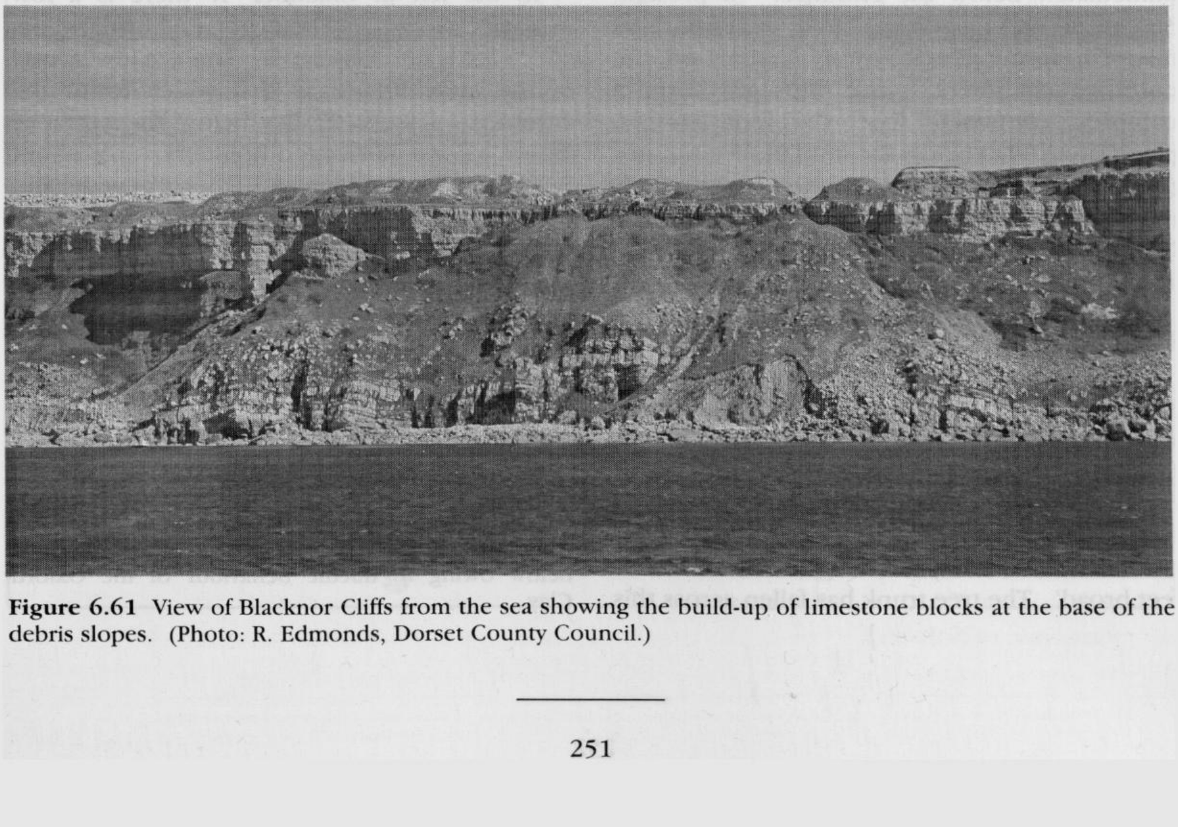


Figure 6.61 View of Blacknor Cliffs from the sea showing the build-up of limestone blocks at the base of the debris slopes. (Photo: R. Edmonds, Dorset County Council.)

BUCKLAND'S WINDYPIT, NORTH YORKSHIRE (SE 588 828)

R.G. Cooper

Introduction

Buckland's Windypit is located on the slope of Far Moor Park, on the right bank of the River Rye in the Duncombe Park Estate, just to the south of Castle Gill, 1 km north of Helmsley, North Yorkshire (see Figure 6.54). The slope at this point is crossed by a complex pattern of hillside trenches up to 2 m deep, leaving an irregular pattern of hummocks clearly visible on stereo-pairs of aerial photographs taken before the present plantation of conifers was begun in the late 1970s. Within this area, and doubtless related to the pattern of trenches, is a fissure in the bedrock, 5 m long and up to 1 m wide, divided into two parts by a large fallen tree trunk. This is the entrance to Buckland's Windypit. The bedrock is Lower Calcareous Grit (Corallian, Upper Jurassic), a fine-grained spicular sandstone conformably overlying Oxford Clay with a slight southerly and easterly dip. The major directions of jointing in this rock are 0°–5° and 95°–100° (Cooper, 1979).

The first recorded descent of the open widened joint was by the Rev. William Buckland, the Professor of Geology at Oxford, in 1822 (Buckland, 1823; Cooper, 1978). However, he penetrated no further than the chambers immediately below the entrance. At present, 366 m of passages have been explored and mapped, forming a labyrinth leading off the entrance chambers. The passages form a complex network on different levels, penetrating 40 m below the level of the entrance (Cooper *et al.*, 1982). Archaeological finds indicating occupation by Bronze Age man have been found in some of the passages (Hayes, 1962, 1987).

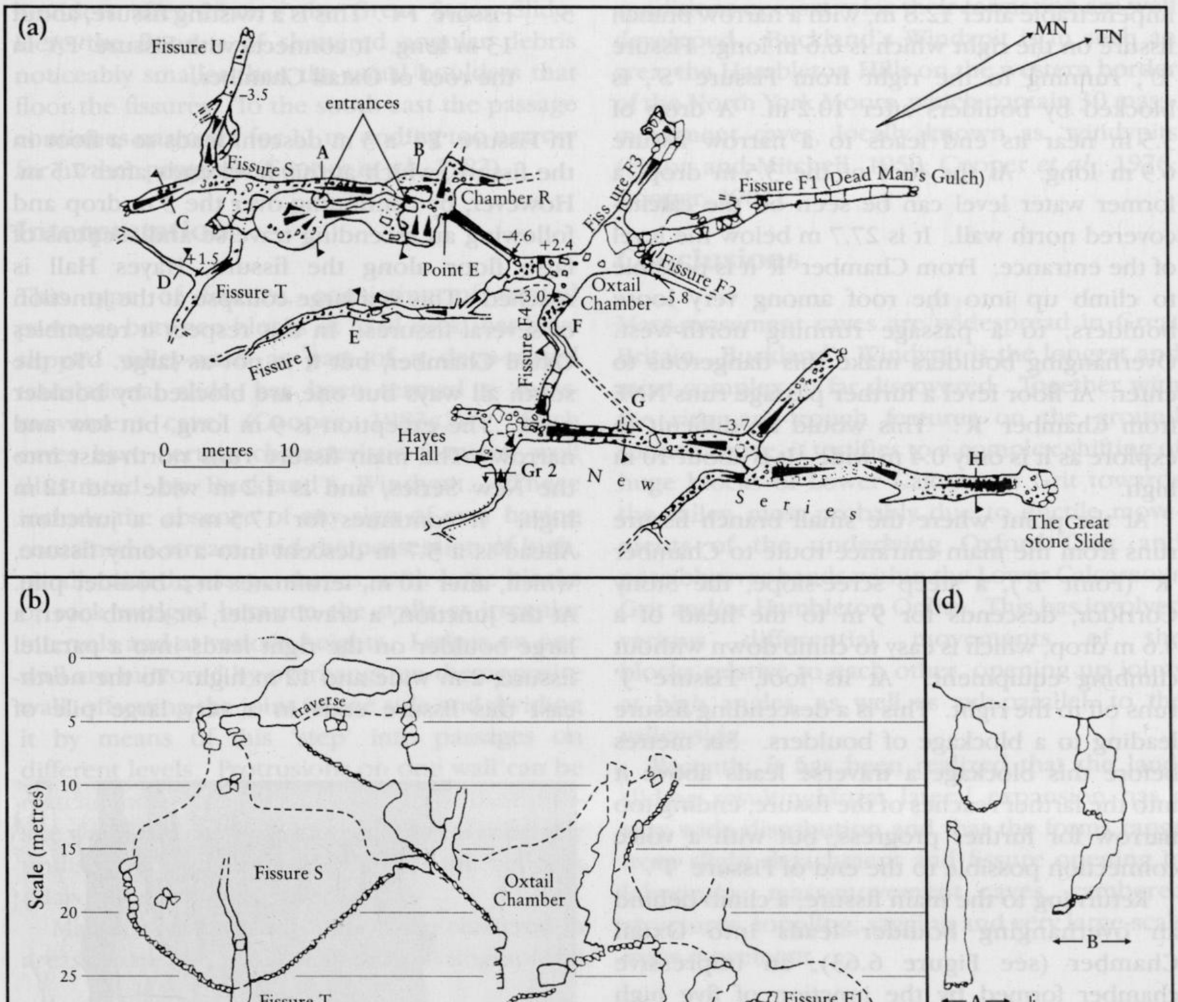
Description

The entrances to this underground labyrinth lie within a wire fence, in a plantation of conifers, a few metres upslope from a forestry track crossing Far Moor Park. Buckland (1823) described the entrance as a 'great irregular crack or chasm…about twenty feet long and three or four feet broad'. The tree trunk has fallen across this fissure, covering it in the middle section and leaving two entrances, one at either end. The labyrinth is large and complex. In the survey (Figure 6.62) and this description, the names given to passages and features are largely those of Fitton and Mitchell (1950) and Hayes (1962). The larger entrance, A, is distinguished by a holly bush growing above it. It requires a climb down of 7.6 m, which needs climbing equipment. However, using the other entrance, B, the entire labyrinth can be explored without climbing equipment. A drop of 2 m lands on a boulder wedged in the top of the fissure. A fixed chain on the right-hand wall can be used to traverse along a ledge leading past a boulder bridge before dropping into a large, light chamber at the foot of entrance A. To the south from this chamber, a short climb and squeeze emerges at the top of Fissure 'S'. Any descent here would involve using climbing equipment in a vertical descent of 22 m among very loose and potentially hazardous boulders wedged in the fissure. This route has been negotiated (Hayes, 1962) but is dangerous.

At the other end of the entrance fissure a descent through boulders leads after 7 m to a vertical drop of 2 m, with an overhang. A climb down to the right-hand side of this gains a ledge skirting under the overhang and following the left wall of the fissure to the head of a steep scree-slope. At this point a small fissure branches off north-west to Chamber 'R', a small chamber at the intersection of several fissures.

To the left in Chamber 'R' there is a low passage that can only be negotiated by crawling, which leads into Fissure 'S'. This is a lower level of the same fissure as that which forms the entrance. Fissure 'S' lies below the entrance fissure's floor of wedged boulders. It is about 20 m long, 13.7 m high and 1.2–2.1 m wide, and is blocked at the far end by boulders. It was the main location at which archaeological material was found in the labyrinth (Hayes, 1987). Two fissures are encountered part of the way along Fissure 'S'. Fissure 'T', on the left, becomes

Figure 6.62▸ Detailed surveys of Buckland's Windypit showing how the lateral spreading of the hillside opens up joint or fissure caves to form a typical labyrinth network. In this case the fissures are beneath the surface suggesting loss of support from below owing to ductile behaviour of the Oxford Clay.


(a)
Fissure U
MN
TN
–3.5
entrances
A
B
Fissure S
C
Chamber R
Fissure F1 (Dead Man's Gulch)
Fissure F3
–4.6
+2.4
+1.5
Point E
D
Fissure T
Fissure F2
Fissure F4
–3.0
Oxtail Chamber
–5.8
E
F
Fissure J
G
0 metres 10
Hayes Hall
3.7
Gr 2
H
N e w
S e r i e s
The Great Stone Slide
(b)
(d)
0
traverse
5
10
15
Fissure S
Oxtail Chamber
Scale (metres)
20
25
Fissure T
Fissure F1
30
B
A
(c)
Dead Man's Gulch
C
D
E
0
5
10
Hayes Hall
F
G
H
15
Fissure S
20
N e w
S e r i e s
25
Fissure J
30
35

impenetrable after 12.8 m, with a narrow branch fissure on the right which is 6.6 m long. Fissure 'U', running to the right from Fissure 'S', is blocked by boulders after 10.2 m. A drop of 3.5 m near its end leads to a narrow fissure 6.9 m long. At the foot of the 3.5 m drop, a former water level can be seen on the calcite-covered north wall. It is 27.7 m below the level of the entrance. From Chamber 'R' it is possible to climb up into the roof among very loose boulders, to a passage running north-west. Overhanging boulders make this dangerous to enter. At floor level a further passage runs NNE from Chamber 'R'. This would be difficult to explore as it is only 0.4 m wide. It is about 10 m high.

At the point where the small branch fissure runs from the main entrance route to Chamber 'R' (Point 'E'), a steep scree-slope, the Stony Corridor, descends for 9 m to the head of a 4.6 m drop, which is easy to climb down without climbing equipment. At its foot, Fissure 'J' runs off to the right. This is a descending fissure leading to a blockage of boulders. Six metres before this blockage a traverse leads above it into the farther reaches of the fissure, ending too narrow for further progress, but with a voice connection possible to the end of Fissure 'T'.

Returning to the main fissure, a climb behind an overhanging boulder leads into Oxtail Chamber (see Figure 6.63), an impressive chamber formed by the junction of five high fissures. These are:

1. The main fissure, by which Oxtail Chamber has been entered.
2. Fissure 'F1', or 'Dead Man's Gulch'. This is reached by a scramble between boulders. A traverse over a hole in the floor, followed by a series of climbable descents from false floors formed from boulders wedged in the fissure, reaches the floor (which could also be founded on wedged boulders). To the south the fissure is choked with boulders below Oxtail Chamber, while to the north a scramble down leads to the deepest point in the labyrinth, 40 m below the entrance.
3. Fissure 'F2'. This is very narrow, but may be descended to a depth of about 9 m with the aid of a ladder, before becoming too narrow.
4. Fissure 'F3'. This involves a scramble up among boulders and ends after 10 m in a loose pile of boulders.
5. Fissure 'F4'. This is a twisting fissure, about 13 m long. It connects with Fissure 'F3' in the roof of Oxtail Chamber.

In Fissure 'F4' a 3 m descent leads to a floor in the fissure, which at this level ends after 7.5 m. However, by continuing over the 3 m drop and following an ascending traverse and sections of false floor along the fissure, Hayes Hall is reached. This is a large collapse at the junction of several fissures. In this respect it resembles Oxtail Chamber, but it is not as large. To the south all ways but one are blocked by boulder piles. The exception is 9 m long, but low and narrow. The main fissure runs north-east into the New Series, and is 1.2 m wide and 12 m high. It continues for 17.5 m to a junction. Ahead is a 3.7 m descent into a roomy fissure, which, after 10 m, terminates in a boulder pile. At the junction, a crawl under, or climb over, a large boulder on the right leads into a parallel fissure, 2 m wide and 12 m high. To the north-east this fissure ends in a very large pile of

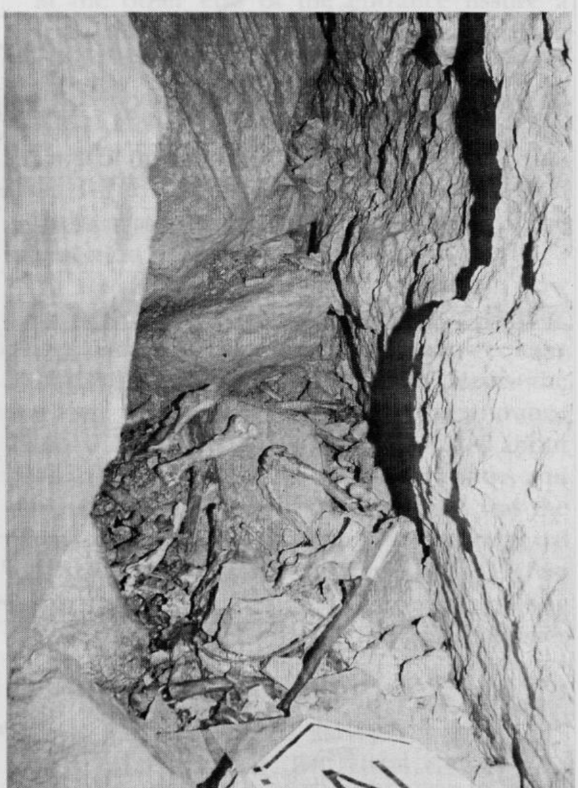


Figure 6.63 Buckland's Windypit showing part of Oxtail Chamber with animal bones. (Photo: © M. Roe.)

boulders after 23 m – the Great Stone Slide. Here the floor is of shattered angular debris noticeably smaller than the usual boulders that floor the fissures. To the south-east the passage continues narrower for 11 m, ending too narrow for further progress (Cooper *et al.*, 1982).

Interpretation

This type of feature, consisting of roofed passages between blocks of hard rock that have slipped valley-ward as part of a deep-seated translational slide, has been termed a 'mass-movement cave' (Cooper, 1983a,b). Such caves have certain characteristic features, well illustrated by Buckland's Windypit. These include the absence of any sign of ever having contained a stream, and the possession of high, parallel-sided passage shapes, with large blocks of rock wedged between the walls at irregular intervals and at various heights. Ledges on one wall are mirrored by overhangs on the opposite wall, offsetting the joint to one side and dividing it by means of this 'step' into passages on different levels. Protrusions on one wall can be matched to corresponding hollows on the opposite wall. The roofs are often formed of relatively undisturbed near-surface layers of the bedrock (Hawkins and Privett, 1981).

Mass-movement caves tend to be clustered in areas where the geological and physiographical conditions necessary for their formation are well developed. Buckland's Windypit is in such an area, the Hambleton Hills on the western border of the North York Moors, which contain 30 mass-movement caves, locally known as 'windypits (Fitton and Mitchell, 1950; Cooper *et al.*, 1976; Cooper, 1981).

Conclusions

Mass-movement caves are widespread in Great Britain. Buckland's Windypit is the longest and most complex so far discovered. Together with the ridge-and-trough features on the ground surface above, it testifies to a complex shifting of huge blocks of Lower Calcareous Grit towards the valley, most probably due to ductile movements of the underlying Oxford Clay, and possibly over bands within the Lower Calcareous Grit and/or Hambleton Oolite. This has involved various differential movements of the blocks relative to each other, opening up joints at high angles, as well as sub-parallel, to the valley-side.

Recently, it has been realized that the land-sliding resulting from lateral expansion has a very wide distribution and that the forms range from slight detachment and fissure opening to labyrinths, mass-movement caves, cambered structures, toppling, sagging and very large-scale mass movement.